



HAL
open science

The role of atmospheric circulation with respect to the interannual variability in the date of snow cover disappearance over northern latitudes between 1988 and 2003

S.M. Vicente-Serrano, Manuela Grippa, Thuy Le Toan, N.M. Mognard

► To cite this version:

S.M. Vicente-Serrano, Manuela Grippa, Thuy Le Toan, N.M. Mognard. The role of atmospheric circulation with respect to the interannual variability in the date of snow cover disappearance over northern latitudes between 1988 and 2003. *Journal of Geophysical Research*, 2007, 112, pp.D08108. hal-00280268

HAL Id: hal-00280268

<https://hal.science/hal-00280268>

Submitted on 13 Aug 2021

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

Copyright

Role of atmospheric circulation with respect to the interannual variability in the date of snow cover disappearance over northern latitudes between 1988 and 2003

Sergio M. Vicente-Serrano,¹ Manuela Grippa,² Thuy Le Toan,² and Nelly Mognard³

Received 6 August 2005; revised 17 October 2006; accepted 16 November 2006; published 26 April 2007.

[1] This paper analyzes the main spatial patterns in the dates of snow cover disappearance variability over northern latitudes between 1988 and 2003. The dates of snow cover disappearance were calculated using satellite passive microwave data from the Special Sensor Microwave/Imager. Spatial and temporal patterns were obtained using principal components analysis in the S mode. We identified eight components, each representing a large region characterized by homogeneous interannual variability in the dates of snow cover disappearance. We found that atmospheric circulation, summarized by means of teleconnection indices, had an important impact on the date of snow cover disappearance for most of these regions. A role is played by the Arctic Oscillation in western Siberia, the spring east Atlantic/west Russian pattern in central Siberia, and the Pacific North American pattern in southern Canada, while the El Niño–Southern Oscillation phenomenon and the west Pacific pattern are significantly related to variability in the date of snow cover disappearance in the northernmost areas of America. For the regions where we found no relationship between the interannual variability in the date of snow cover disappearance and the teleconnection indices, a direct relationship with some spatial patterns of sea level pressures, which are not well summarized by the teleconnection indices, was found.

Citation: Vicente-Serrano, S. M., M. Grippa, T. Le Toan, and N. Mognard (2007), Role of atmospheric circulation with respect to the interannual variability in the date of snow cover disappearance over northern latitudes between 1988 and 2003, *J. Geophys. Res.*, *112*, D08108, doi:10.1029/2005JD006571.

1. Introduction

[2] Snow plays a major role in different physical, ecological and biological processes affecting land surface temperature and atmospheric circulation through albedo and soil moisture [Groisman *et al.*, 1994; Barnett *et al.*, 1989; Yasunari *et al.*, 1991]. The spatial and temporal variability of snow cover (quantified by means of different variables such as the extent and duration of snow cover, snowmelt onset and timing, snow-off date, snow cover disappearance date, end date of continuous snow cover, etc.) has a major impact on river runoff and flows at northern latitudes [Fukutomi *et al.*, 2003; Quinton *et al.*, 2004; Yang *et al.*, 2003, 2004; Ye *et al.*, 2003] and can have a significant influence on the salinity and sea ice distribution of the Arctic Ocean [Aagaard and Carmack, 1989; Peterson *et al.*, 2002], thus affecting the atmospheric circulation of the entire Northern Hemisphere [Thompson and Wallace, 1998]. Several studies have highlighted the impact of snowmelt

timing on climate by means of ocean temperature changes, surface albedo modifications and soil moisture after melting [Cohen and Rind, 1991; Groisman *et al.*, 1994; Saunders *et al.*, 2003; Saito and Cohen, 2003; Gong *et al.*, 2004; Saito *et al.*, 2004]. Sankar-Rao *et al.* [1996] showed that the Eurasian snow cover and melting period play a significant role in the development of the Asian summer monsoon.

[3] Snowmelt is also very important for ecosystems and biological processes. It influences both plant phenology [Galen and Stanton, 1995; Ouerbauer *et al.*, 1998; Van Wijk *et al.*, 2003; Dye and Tucker, 2003; Grippa *et al.*, 2005a] and the carbon cycle via the increased ecosystem respiration after melting [Goulden *et al.*, 1998; Chapin *et al.*, 2000; Frauenfeld *et al.*, 2004]. Moreover, snow also plays a role in maintaining carbon fluxes during the winter through its insulating effect on soil temperatures [McDowell *et al.*, 2000].

[4] Recent observations have shown that snowfall and snow depth during the cold season in the Northern Hemisphere have increased from 3.9 to 5.9% over the past few decades [Brown *et al.*, 1995; Ye *et al.*, 1998; Groisman and Easterling, 1994; Kitaev *et al.*, 2002] as a result of the positive precipitation trends at northern latitudes [Ye and Mather, 1997; Schönwiese and Rapp, 1997; Heino *et al.*, 1999]. However, the snow-free period has also increased

¹Instituto Pirenaico de Ecología, Spanish Research Council, Zaragoza, Spain.

²Centre d'Etudes Spatiales de la Biosphère, Toulouse, France.

³Laboratoire d'Etudes en Géophysique et Océanographie Spatiales, Toulouse, France.

because the snowmelt date has advanced 9–15 days in the last 30 years [Dye, 2002; Zhou *et al.*, 2001; Serreze *et al.*, 2000; Stone *et al.*, 2002; Bamzai, 2003] due to the temperature increase observed at northern latitudes [Jones and Moberg, 2003]. In contrast, no particular trends have been found for the dates of snow onset [Bamzai, 2003; Dye, 2002], although Brown [2000] did show a widespread significant trend toward earlier dates of maximum snow depth over the 1955–1997 period in North America.

[5] It is therefore extremely important to improve our understanding of the physical mechanisms regulating the spatial and temporal variability of snowmelt timing on a global scale. Several authors have linked this to air temperature variability [Foster *et al.*, 1983; Karl *et al.*, 1993; Nijssen *et al.*, 2001; Regonda *et al.*, 2005]. Brown [2000] showed a greater sensitivity of snow extent to temperatures in November and April over large areas of the Northern Hemisphere. This is important because higher temperatures in November would delay the snow fall in autumn and higher temperatures in April would produce an earlier snowmelt in spring. Simple relationships between temperature and snow cover disappearance can be derived for some areas, usually large continental regions with clear skies and shallow snow cover such as prairie and steppe where temperature and snow cover are closely linked [Ohmura, 2001; Hock, 2003; Walter *et al.*, 2005], but not for those where nontemperature related processes are more important for the snowmelt process, for example, albedo, long wave radiation, atmospheric transmissivity, ground heat conduction, wind speed, etc. [Fontaine *et al.*, 2002; Rango and Martinec, 1995; Albert and Krajewski, 1998; Essery and Etchevers, 2004].

[6] An interesting way of investigating the spatial patterns and temporal variability in snowmelt timing is to study their relationship to atmospheric circulation indices that summarize the behaviour of different climate elements such as precipitation, temperatures, cloudiness, wind speed, etc. [Trenberth, 1990; Ropelewski and Halpert, 1987; Thompson *et al.*, 2002; Marshall *et al.*, 2001]. The use of these indices, rather than complex models requiring several inputs, may simplify the prediction of snowmelt timings.

[7] Several studies have investigated the links between atmospheric circulation patterns and snow cover and, in some cases, found significant relationships that would be useful for seasonal forecasting [e.g., Saunders *et al.*, 2003; Gutzler and Rosen, 1992; Clark *et al.*, 1999; Morinaga *et al.*, 2003; Bamzai, 2003].

[8] However, the relationship between atmospheric circulation and the date of snow cover disappearance in the whole of the Northern Hemisphere has yet to receive much attention. Gutzler and Rosen [1992] compared the evolution of two major atmospheric circulation indices, the Pacific North American pattern and the North Atlantic Oscillation, while Bamzai [2003] analyzed the role of the Arctic Oscillation. However, most studies on the spatial and temporal variability of snow cover in the Northern Hemisphere have focused on trends analysis [Brown, 2000; Dye, 2002; Smith and Saatchi, 2004], and those investigating the relationships with atmospheric circulation have mainly been carried out at the continent or country level [e.g., Bednorz, 2002; Kryjov, 2002; Schaefer *et al.*, 2004; Hori and Yasunari, 2003].

[9] This paper analyses the spatial and temporal variability in the date of snow cover disappearance at northern latitudes over the whole Northern Hemisphere by means of remote sensing data. The objective was to understand how the general atmospheric circulation, summarized by means of teleconnection indices, can determine the spatial and temporal variability in the date of snow cover disappearance. Our analysis enables us to locate the areas where atmospheric circulation and/or temperature alone can be used for modeling or prediction purposes. This approach can be very useful for quantifying the possible effects of climate change in relation to modifications of atmospheric circulation.

2. Database and Methods

2.1. Snowmelt Timing Calculation by Means of SSM/I Data

[10] Passive microwave measurements contain information on both snow extent and snow depth, regardless of solar illumination and cloud cover, and are therefore better than optical sensors for monitoring high latitude areas that are often covered by clouds [Grippa *et al.*, 2004].

[11] In this study we use the passive microwave radiances measured by the Special Sensor Microwave/Imager (SSM/I), on board the Defence Meteorological Satellite Program satellite series since July 1987 (DMSP F-8, F-11 and F-13 platforms). SSM/I is a multispectral radiometer with horizontally and vertically polarized channels at 19, 37 and 85.5 GHz and a vertically polarized channel at 22.235 GHz. This study employs the spectral gradient, which is the difference between the horizontally polarized 19 GHz and 37 GHz brightness temperatures. The radiometer sensitivity or noise equivalent temperature differential is 0.41 K in the 19 GHz and 0.37 K in the 37 GHz horizontally polarized channels.

[12] The National Snow and Ice Data Center (NSIDC) provided the SSM/I data mapped to the Equal Area SSM/I Earth Grid, with a 25 km² resolution. To minimize the spatial gaps resulting from the swath width the daily data were averaged over pentads (5-day averages). The morning satellite passes only were selected for this analysis.

[13] The 37 GHz radiance is scattered by snow grains and the brightness temperature decreases as the snow depth increases, while the 19 GHz horizontally polarized channel is sensitive to the effects of ground temperature and atmospheric perturbations.

[14] When snow is absent the mean value of the spectral gradient is about 3 K. The date of snow cover disappearance was therefore calculated as the date at which the spectral gradient was less than 3 K for at least three consecutive pentads (Figure 1). The microwave brightness temperatures are attenuated by scattering with trees in densely forested areas. Different empirical formulations have been proposed to take this into account by including the forest fractional cover into snow retrieval algorithms [Pulliainen and Hallikainen, 2001; Kelly and Chang, 2003].

[15] In this study we did not apply any corrections for this effect. Although this may bias the absolute values of the snow estimates derived over dense coniferous forests (later snow cover disappearance), it does not noticeably affect their interannual variability and therefore it has no impact

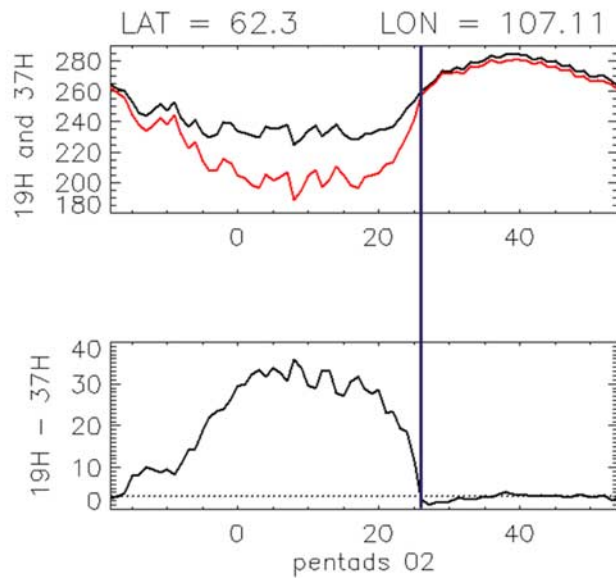


Figure 1. (top) SSM/I measured brightness temperatures (K) at a given location during winter 2002. (bottom) Spectral gradient and determination of the snowmelt date according to the 3K threshold.

on our analysis. In fact, we carefully checked the brightness temperature values over different land cover types, and even if they are greatly attenuated over dense forest the seasonal evolution of the spectral gradient is sensitive enough to discriminate between snow and no snow, and therefore to determine the interannual variations in the timings of snow cover disappearance.

[16] Moreover, the interannual variability in the date of snow cover disappearance derived by this method has been validated with good results through comparison with river discharge measurements in the Ob river basin, which contains a wide range of vegetation types [Grippa *et al.*, 2005b]. Armstrong and Brodzik [2001] compared the snow extent provided by NOAA weekly charts derived from visible-band satellite imagery and the snow extent derived by passive microwave sensors and reported a good agreement between the two data sets during the spring season. The advantage of using passive microwaves over optical sensors is that, due to their greater penetration, they are less affected by cloud and land cover. However, they are less sensitive to very shallow snowpacks and their potential is limited over mountainous areas since brightness temperatures are affected by topographic changes.

2.2. Principal Components Analysis of Variability in the Date of Snow Cover Disappearance

[17] To categorize the interannual spatial variability in the date of snow cover disappearance we used principal components analysis (PCA). PCA reduces a large number of interrelated variables to a few independent principal components that capture much of the variance of the original data set [Hair *et al.*, 1998]. It also generates a matrix of loading values, which can be displayed graphically, showing the spatially coherent modes of variability.

[18] The application of PCA to different time series in a specific area can be performed in T or S modes [Jolliffe, 1990; Serrano *et al.*, 1999]. The S mode is used to determine the main temporal patterns of the time series evolution. In this paper we used PCA in the S mode from the time series of each pixel to identify the main modes of variability. To retain a consistent signal and prevent snow cover changes from altering the analysis, we only considered areas with consistent snow data during the whole study period (1988–2003). Therefore this study only addresses the northernmost areas of the Northern Hemisphere. This restriction is necessary as the application of PCA requires complete temporal series.

[19] A total of 16,607 pixels of 0.5° degrees were analyzed. The series of dates of snow cover disappearance at each grid point were standardized according to the average and the standard deviation over the period 1988–2003. A correlation matrix providing an efficient representation of the data set variance was selected [Barry and Carleton, 2001].

[20] The number of components that explain a representative percentage of the variance was selected according to a threshold of around 70% of total variance [Briffa *et al.*, 1994]. In addition, we considered a subjective criterion according to the distance among component error intervals [Cattell, 1966; Esteban *et al.*, 2005], and also took into account the spatial representativeness of components.

[21] The rotation simplifies the spatial patterns of the variable studied [Barnston and Livezey, 1987], redistributes the final explained variance, thus enabling a clearer separation of components while maintaining their orthogonality [Hair *et al.*, 1998], and concentrates the loading for each PC onto the most influential variables. We used the Varimax rotation [Kaiser, 1958], which is the most widely applied option as it produces more stable and physically robust patterns [Richman, 1986; White *et al.*, 1991; Ye, 2000]. The representative areas of each component were determined by mapping the rotated empirical orthogonal functions (REOFs), which represent the parametric correlation between the components and the series of the date of snow cover disappearance in each pixel.

2.3. Atmospheric Circulation Patterns in the Northern Hemisphere

[22] The principal patterns of atmospheric circulation variability have been analyzed and described in numerous studies [Rogers and Van Loon, 1979; Wallace and Gutzler, 1981; Mo and Livezey, 1986; Barnston and Livezey, 1987]. These atmospheric circulation patterns affect the climate of large areas of the Northern Hemisphere [e.g., Trenberth, 1990; Zorita *et al.*, 1992; Hurrell, 1995; Hurrell and Van Loon, 1997; Gershunov and Barnett, 1998].

[23] Mo and Livezey [1986] showed teleconnections in sea level pressures (SLP) between regions separated by thousands of kilometers. The simultaneous variations in the climatic system are recurrent and produce SLP patterns known as teleconnections. Barnston and Livezey [1987] summarized the main atmospheric circulation patterns in the Northern Hemisphere using teleconnection indices. In this paper we use the teleconnection indices developed by the Climate Prediction Center of the United States (<http://www.cpc.ncep.noaa.gov/data/teledoc/telecontents.shtml>).

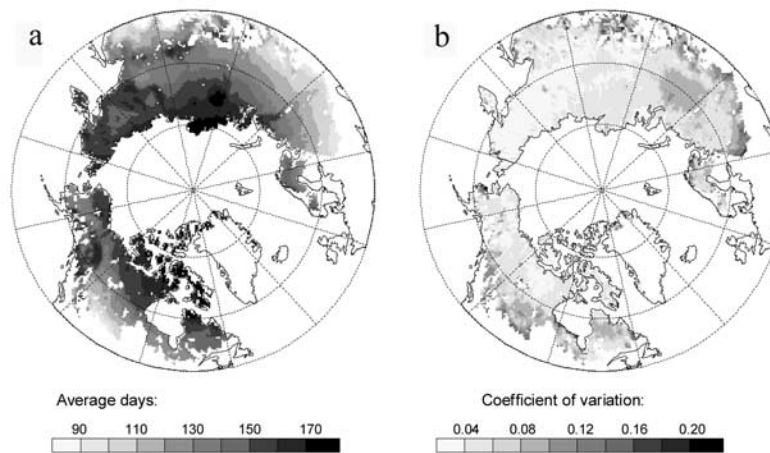


Figure 2. (a) Spatial distribution of the average date of snow cover disappearance and (b) the coefficient of variation of the date of snow cover disappearance between 1988 and 2003.

These are calculated through monthly mean standardized 500-hPa height anomalies using rotated principal components analysis (RPCA), following *Barnston and Livezey* [1987]. The indices considered for this paper are the North Atlantic Oscillation (NAO), the east Atlantic (EA) pattern, the East Atlantic Jet (EA JET), the east Atlantic/western Russia (EA/WR) pattern, the Scandinavian (SCA) pattern, the west Pacific (WP) pattern, the North Pacific (NP) pattern and the Pacific North American (PNA) pattern. These indices appear during most months and explain the majority of atmospheric circulation variability in the Northern Hemisphere (60–80%, with some seasonal differences). A detailed explanation of these circulation patterns is given by *Barnston and Livezey* [1987], Climate Prediction Center (<http://www.cpc.ncep.noaa.gov/data/teledoc/telecontents.shtml>), and *Washington et al.* [2000].

[24] In addition, we also used two global atmospheric indices that affect the whole terrestrial climate system: El Niño–Southern Oscillation (ENSO) and the Arctic Oscillation (AO). The ENSO comprises the main factors in climatic variability on a global scale [*Trenberth*, 1997; *Kiladis and Diaz*, 1989; *Halpert and Ropelewski*, 1992] and has an important influence on the climate of the Northern Hemisphere [*Diaz and Kiladis*, 1992; *Halpert and Ropelewski*, 1992; *Ropelewski and Halpert*, 1987; *Fraedrich*, 1994]. This atmosphere–ocean coupled mode is quantified by means of the Southern Oscillation index (SOI), which is calculated as the normalized difference between the pressure anomalies recorded at the observatories of Tahiti (east Pacific) and Darwin (west Pacific) [*Ropelewski and Jones*, 1987]. The SOI was obtained from Climate Research Unit of the University of East Anglia (<http://www.cru.uea.ac.uk/cru/data/soi.htm>).

[25] The AO is calculated as the first component from the PCA of 1000 hPa height data in the Northern Hemisphere [*Thompson and Wallace*, 1998]. Its variability is caused by an ocean–tropospheric–stratospheric connection [*Thompson et al.*, 2002]. *Baldwin and Dunkerton* [1999] indicated that AO is the main tropospheric–stratospheric variability system in the Northern Hemisphere and that it affects significantly the climate of large areas [*Thompson et al.*, 2002; *Kerr*, 2001]. The AO index used here was

obtained from the Climate Prediction Center, United States (http://www.cpc.ncep.noaa.gov/products/precip/CWlink/daily_ao_index/ao.shtml).

[26] The role of the atmospheric circulation patterns described above with respect to the interannual variability in the date of snow cover disappearance was studied by means of parametric correlations (R-Pearson) between teleconnection indices and PCA components. All possible seasonal and monthly relationships were analyzed, and temporal lags were also considered.

[27] To investigate the role of temperatures with respect to the variability in the date of snow cover disappearance, we used the monthly surface air temperatures at 0.5° (TS 2.0 data set), as developed by the Climate Research Unit of the University of East Anglia [*Mitchell et al.*, 2003]. Temperature averages corresponding to each component were calculated over the area where REOF loadings were statistically significant ($p < 0.01$).

[28] Finally, we analyzed the relationship between teleconnection indices and SLP values by means of parametric correlations, using the Northern Hemisphere monthly SLP grids for the period 1988 to 2003, provided by NCEP–NCAR (<http://dss.ucar.edu/datasets/ds010.1/>) [*Trenberth and Paolino*, 1980; *Basnett and Parker*, 1997].

3. Results and Discussion

3.1. Average Distribution of the Date of Snow Cover Disappearance and Its Variability in the Northern Hemisphere

[29] Figure 2 shows the spatial distribution of the average date of snow cover disappearance and the coefficient of variation between 1988 and 2003. The range oscillates between day 90 (end of March) in the south of Eurasia and North America and day 170 (beginning of July) in northern Siberia. Latitude seems to control the general patterns for the average date of snow cover disappearance, this being related to the spatial distribution of temperatures in agreement with the results obtained using AVHRR data [*Dye*, 2002]. The spatial differences in the temporal variability in the date of snow cover disappearance (Figure 2b) show a high interannual variability in the southern areas of

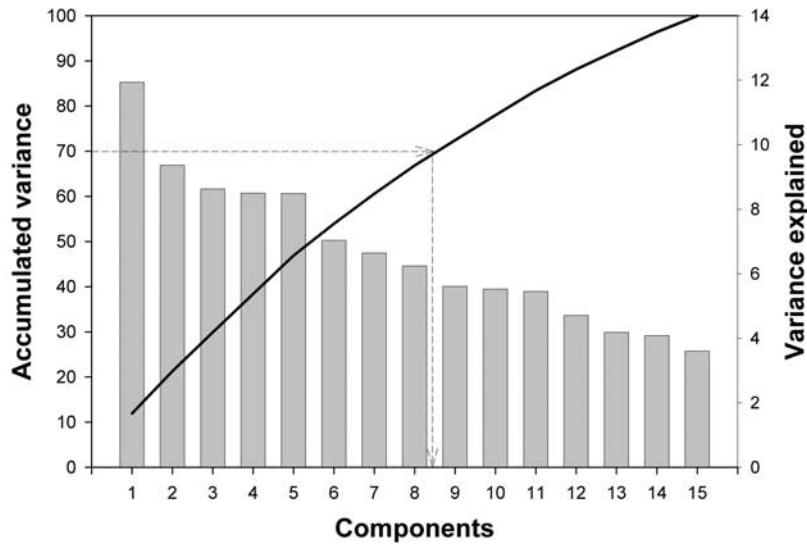


Figure 3. Percentage of variance explained by each component obtained from rotated principal component analysis in S mode. Date of snow cover disappearance (1988–2003).

North America, Eurasia, Scandinavia and western Siberia (with values higher than 0.10) and a low interannual variability in the north and northeast of Siberia and in some areas of northern Canada.

3.2. Spatial Patterns of Temporal Variability in the Date of Snow Cover Disappearance

[30] Figure 3 shows the results of the S mode PCA. The first eight components explain more than 70% of the temporal variance in the date of snow cover disappearance in the Northern Hemisphere and they were therefore selected for further analysis.

[31] Figures 4a and 4b show the spatial distribution of the REOFs for the first eight components. Each of them individualizes areas of homogeneous interannual variability in the date of snow cover disappearance.

[32] There are no significant trends in the temporal series for any of the components (Rho-Spearman test, $p < 0.05$). This is in agreement with the results reported by *Dye* [2002] using AVHRR data over the two last decades of the twentieth century, and those of *Smith and Saatchi* [2004] using SSM/I data. However, several analyses over longer time periods than the sixteen years considered here have shown significant trends toward an earlier date of snow

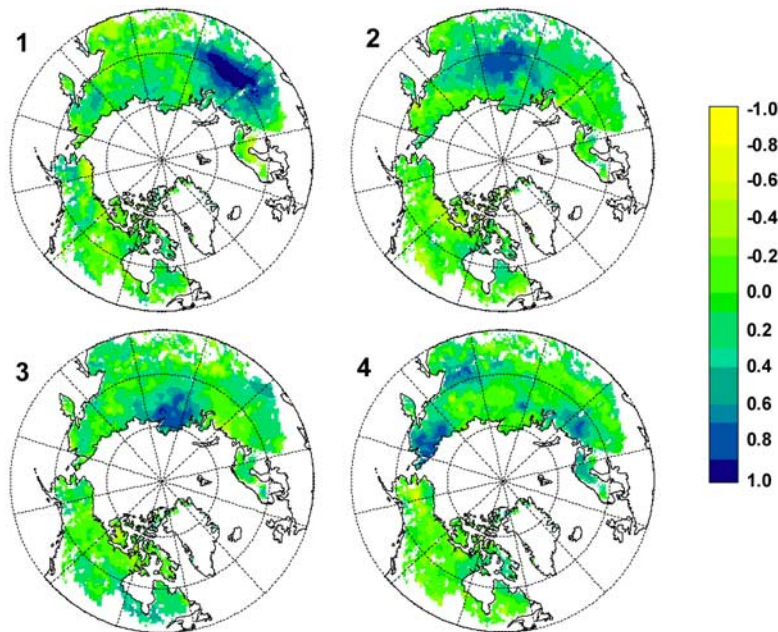


Figure 4a. Components 1–4 rotated empirical orthogonal functions (REOF) obtained from PCA (date of snow cover disappearance, 1988–2003).

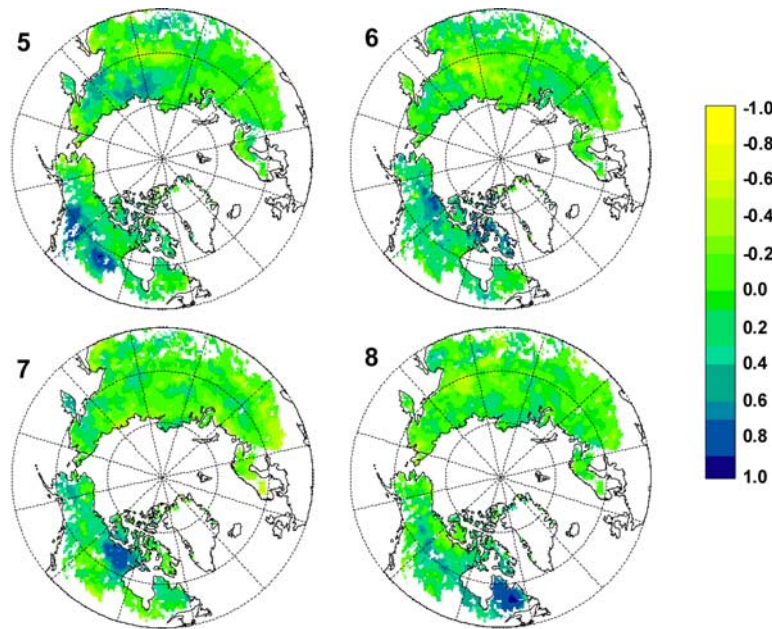


Figure 4b. Components 5–7 rotated empirical orthogonal functions (REOF) obtained from PCA (date of snow cover disappearance, 1988–2003).

cover disappearance and a longer snow-free period in the Northern Hemisphere [Brown, 2000; Robinson et al., 1995; Dye, 2002; Laxon et al., 2003; Stone et al., 2002; Bamzai, 2003].

3.3. Role of Atmospheric Circulation Patterns With Respect to Variability in the Date of Snow Cover Disappearance in Each Region

[33] We found significant correlations between some of the teleconnection indices and the components representing the variability in the date of snow cover disappearance in different regions. These correlations are summarized in Table 1. Figure 5 shows the temporal evolution of each of these components together with the temporal evolution of any significantly correlated teleconnection index. Figure 6 shows areas of significant correlations (positive and negative) between the date of snow cover disappearance and the teleconnection indices reported in Table 1.

[34] Table 1 also reports the values of the significant correlations between the temporal series of each component and the average temperature in the corresponding areas recorded during the snowmelt period. Correlations were

always negative, higher temperatures being related to an earlier date of snow cover disappearance.

3.3.1. Component 1: Western Siberia

[35] There is a significant correlation between component 1 and the previous summer Arctic Oscillation (AO) and North Atlantic Oscillation (NAO). The highest correlation was found for the month of June ($R = -0.68$ for the AO and $R = -0.62$ for the NAO, both significant at 99%). The AO and the NAO are strongly related [Ambaum et al., 2001] and the NAO can be considered as a regional pattern of the more general AO pattern [Thompson and Wallace, 1998; Wallace, 2000], which controls the climate in large regions of Eurasia, the North Atlantic and West Africa [Lamb and Pepler, 1987; Hurrell and Van Loon, 1997]. The spatial distribution of correlations between the AO index in June of the previous year and the date of snow cover disappearance (Figure 6) shows significant negative coefficients over large areas of western Siberia that coincide with the areas represented by component 1. Average temperatures over the regions individuated by component 1 are also significantly correlated with the AO (the correlation between mean April–May temperature and the previous June AO index is $R = 0.68$ ($p < 0.01$)).

Table 1. Significant Correlations Found Between the Teleconnection Indices, Average Temperatures, and the Components Obtained From the Series of Date of Snow Cover Disappearance

Component	Teleconnection Index	Period of the Teleconnection Index	Correlation (R)	Correlation (R) Temperature
Component 1	AO	June previous year	$R = -0.68$ ($p < 0.01$)	April–May, $R = -0.80$ ($p < 0.01$)
Component 2	SOI	summer previous year	$R = -0.66$ ($p < 0.01$)	April–May, $R = -0.84$ ($p < 0.01$)
Component 2	EA-WR	spring current year	$R = -0.56$ ($p < 0.05$)	April–May, $R = -0.84$ ($p < 0.01$)
Component 3	None			June, $R = -0.67$ ($p < 0.01$)
Component 4	None			
Component 5	PNA	May current year	$R = -0.61$ ($p < 0.01$)	May, $R = -0.56$, ($p < 0.05$)
Component 6	SOI	spring current year	$R = 0.66$ ($p < 0.01$)	
Component 6	WP	spring current year	$R = -0.84$ ($p < 0.01$)	
Component 7	AO	winter current year	$R = 0.69$ ($p < 0.01$)	June, $R = -0.63$, ($p < 0.05$)
Component 8	none			April–May, $R = -0.63$, ($p < 0.05$)

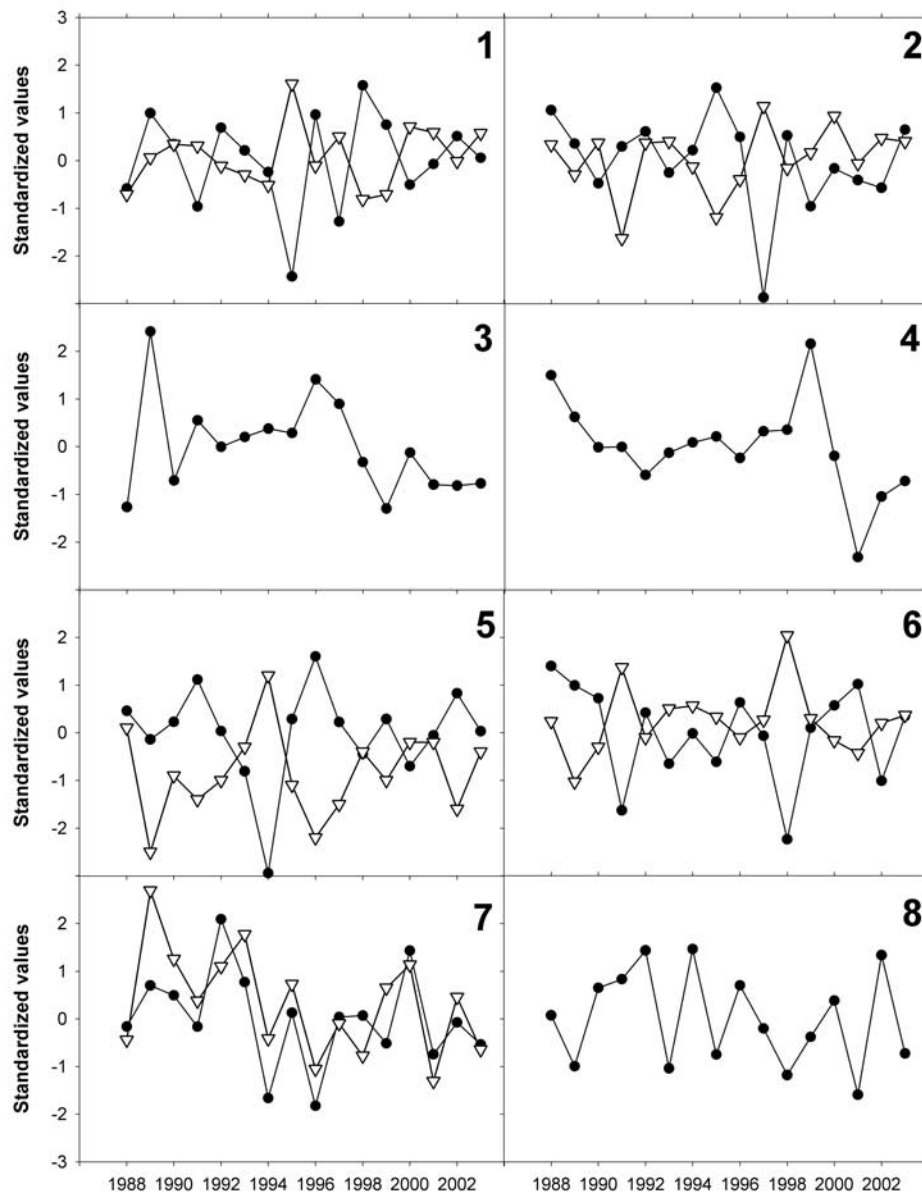


Figure 5. Evolution of the eight components obtained from the series of the date of snow cover disappearance (circles). The triangles represent the following: component 1, AO (June of the previous year); 2, EA-WR (spring of the current year); 5, PNA (May of the current year); 6, WP (spring of the current year); 7, AO (Winter of the current year).

[36] Significant lag correlations between climate elements and the ocean and/or atmospheric characteristics in western Siberia have been emphasized in previous studies. *Bamzai* [2003], using NOAA-NESDIS weekly snow cover observations, showed that the winter AO index affects negatively and significantly the snow cover for the seven months between January and July. *Ogi et al.* [2004] found a significant correlation between the summer and winter AO, which could explain the 10-month lag we observed between the June AO and the date of snow cover disappearance in the following year.

[37] *Ye* [2001] conducted a PCA from winter precipitation in northern Eurasia and found a similar spatial pattern to that shown here. He also found a strong correlation

between winter precipitation and the sea surface temperature (SST) of the Atlantic Ocean during the previous winter (1-year lag). The spatial distribution of this correlation is highly related to a pattern of atmospheric circulation that resembles the AO/NAO pattern. Several studies have shown a close relationship between atmospheric circulation in the Northern Hemisphere and SST anomalies over the Pacific and Atlantic oceans [*Häkkinen*, 2000; *Lau*, 1997; *Corti et al.*, 2000]. *Livezey and Smith* [1999] indicated that an atmospheric circulation pattern similar to the NAO configuration is closely related to spatial patterns of SST anomalies in the Atlantic. *Ye* [2001a] provided evidence that the same spatial pattern of SST anomalies affects winter precipitation in western Siberia with a 1-year lag.

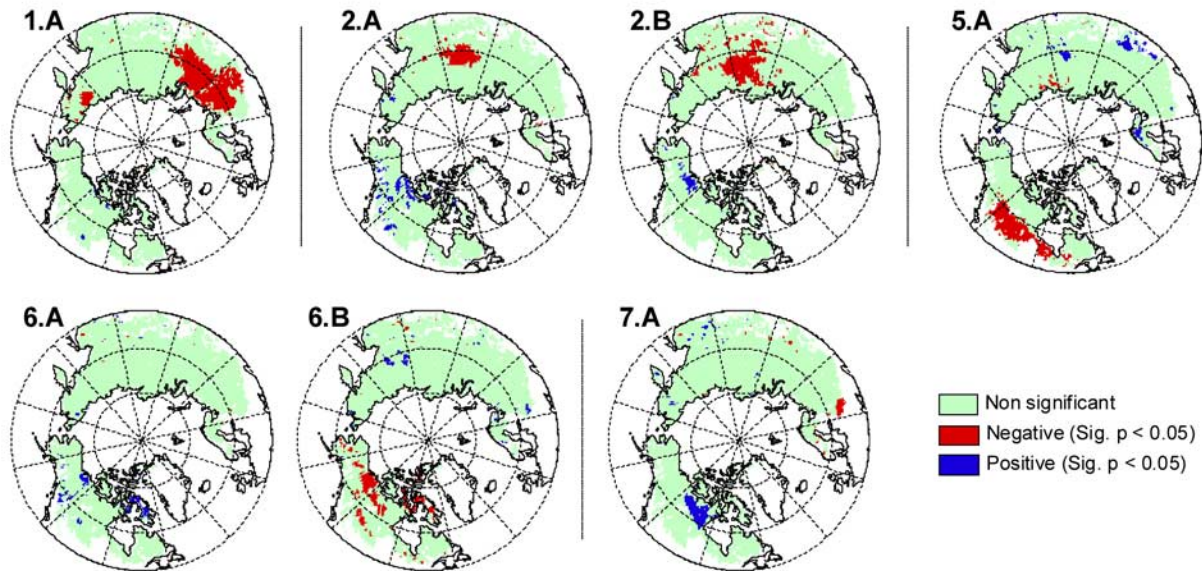


Figure 6. Spatial distribution of correlations between the standardized dates of snow cover disappearance and the teleconnection indices indicated in the Table 1. The numbers of the different charts indicate the components in which significant correlations have been found to the corresponding teleconnection indices. Components: 1.A, AO (June of the previous year); 2.A, SOI (summer of the previous year); 2.B, EA-WR (spring of the current year); 5.A, PNA (May of the current year); 6.A, SOI (spring of the current year); 6.B, WP (spring of the current year); 7.A, AO (winter of the current year).

3.3.2. Component 2: Central Siberia

[38] There is a significant correlation ($R = -0.56$, $p < 0.01$) between component 2 and the average spring (March, April and May) east Atlantic/west Russian (EA/WR) pattern. Positive phases of the EA/WR pattern are characterized by negative pressure/height anomalies throughout western and southwestern Russia and positive pressure/height anomalies over northwestern Europe and the south of Asia [Barnston and Livezey, 1987; Krichak and Alpert, 2005]; southwest flows are caused in central Siberia. The negative phase is related to the contrary pressure/height anomalies and to dominant northern flows. Therefore the positive phase is related to early snowmelt while the northern flows during negative phases are related to cloudy and cold conditions, which can retard the snowmelt onset and increase the duration of the freezing period. The spatial distribution of correlations between the date of snow cover disappearance and the spring EA/WR pattern (Figure 6) shows areas of negative correlations that correspond to the areas represented by component 2.

[39] These results agree with those reported previously. Morinaga *et al.* [2003] found a significant and negative correlation between snow depth and the EA/WR pattern in January ($R = -0.47$). Ye [2001b] analyzed the interannual variability of the winter snow accumulation in northern Eurasia between 1936 and 1995 using rotated PCA and showed a snowpack pattern representative of central Siberia and western Russia, which is connected to the EA/WR pattern. Clark *et al.* [1999] found the EA/WR pattern to be the major mechanism affecting winter snow cover in this region and showed that during negative EA/WR phases the snow cover area is larger than normal due to lower temperatures. They in fact found that negative/positive EA/WR

extremes cause negative/positive temperature anomalies (greater than 2°C) in central Siberia. Negative phases also correspond to a marked extension of the Siberian high, consistent with both the anticyclonic vorticity advection and reduced air humidity. Positive phases in the EA/WR pattern are associated with southwesterly flows, which Ueda *et al.* [2003] related to incoming warm air in this region and which could be responsible for the rapid snowmelt in March–April.

[40] Moreover, in the region represented by component 2 a significant correlation between the SOI of the previous summer and the variability in the date of snow cover disappearance has been found. For central Siberia, Vicente-Serrano *et al.* [2006a] found a significant connection between the spring phenology of boreal forests and the previous summer SOI between 1982 and 2004. This would be due to the effect of the SOI on sea level pressures, 500 hPa geopotential and the wind flow direction and intensity during April and May, and noticeably affects the average temperatures during these months, with a positive correlation between the April–May surface temperatures and the SOI of the previous summer.

3.3.3. Component 3: Northern Siberia

[41] In northern Siberia (component 3) no significant correlations were found with any of the atmospheric circulation patterns analyzed, although a significant correlation was found with temperatures in June ($R = -0.67$, $p < 0.01$), the month in which snow cover generally disappears in this area.

3.3.4. Component 4: Eastern Siberia/Eastern Europe

[42] Component 4 shows a higher spatial heterogeneity than the previous one, representing areas of both eastern and western Siberia. No significant correlations were found with

any teleconnection index or with the average temperature of these regions.

3.3.5. Component 5: Southern Canada

[43] Component 5, representing southern Canada, is significantly correlated with the spring (March, April, and May) Pacific/North American (PNA) pattern ($R = -0.61$, $p < 0.01$), although the correlation is higher when considering the May PNA index ($R = -0.74$, $p < 0.01$). The PNA pattern is one of the most prominent modes of low-frequency variability in the Northern Hemisphere [Barnston and Livezey, 1987; Sheng et al., 1998]. The positive phase is characterized by negative pressures/low anomalies in both the Gulf of Alaska and Mexico and by positive pressures/high anomalies in central Canada. The positive phase shows dominant high pressures over the areas represented by component 5. The spatial distribution of correlations between the spring PNA and the variability in the date of snow cover disappearance shown in Figure 6 (component 5.A) is consistent with the areas represented by component 5. The effect of the PNA on the climate of North America has been identified by numerous authors. Redmond and Koch [1991] showed the PNA to have a significant role in explaining the surface climate and streamflow variability in the northwest United States between October and March. Furthermore, Gutzler and Rosen [1992] analyzed the PNA pattern in order to explain the winter variability in snow cover in the whole Northern Hemisphere and found a significant negative correlation over North America, mainly at longitudes between 170°W and 105°W .

3.3.6. Component 6: Northeastern Canada and the Northern Islands

[44] Component 6, which represents the northernmost areas of North America, is significantly correlated with the spring Southern Oscillation index ($r = 0.66$, $p < 0.01$) and the spring west Pacific (WP) pattern ($r = -0.84$, $p < 0.01$). In addition to the correlations with the autumn SOI, we also found significant correlations with the SOI of the previous year (since the autumn). This is due to the low monthly and seasonal temporal inertia of this ocean-atmosphere coupled mode, which shows a high autocorrelation between monthly values for periods between two and twelve months [Trenberth, 1997].

[45] There are significant positive correlations between the spring SOI and the date of snow cover disappearance in region 6. Although the opposite sign is found for the WP pattern, the spatial patterns of correlation coefficients are similar. Several studies have highlighted the important role of the ENSO in the climate of northern Canada and Alaska, and with respect to the connection between the southern and northern Pacific Ocean [Gershunov and Barnett, 1998; Lluch-Cota et al., 2001]. Mantua et al. [1997] found that the principal mode of snowpack variability over the western United States shows a strong association with the North Pacific atmospheric circulation by means of a pattern that resembles the WP pattern and which is related to ENSO. Moreover, Carrera et al. [2004] found significant high-pressure blocking events corresponding to El Niño phases in the northeast Pacific. Trenberth and Hurrell [1994] related El Niño to the warming in northwest America through the northward advection of warmer air from the North Pacific. Niebauer [1998] analyzed the role of the SOI and the WP pattern in the ice cover in the Bering Sea. He

showed that El Niño causes the Aleutian low to move eastward of normal, carrying warm Pacific air over the Bering Sea, while La Niña is associated with larger pressure moving over the Bering Sea and that these phenomena have important effects on the ice cover of the Bering Sea. These processes and climate connections were also observed using instrumental data. Temperatures are significantly warmer than normal in east Alaska during El Niño years, while La Niña winters produce temperatures significantly below normal [Papineau, 2001]. These findings enabled a connection to be made between the ENSO and snow cover in some regions of North America [Gutzler and Preston, 1997; Smith and O'Brien, 2001].

3.3.7. Component 7: Northern Canada

[46] Component 7 also represents a homogeneous region in the north of Canada. In this area the date of snow cover disappearance is generally recorded in June. There is a significant correlation with the average temperature recorded during this month ($R = -0.63$, $p < 0.01$) and also a significant positive correlation with the December-January-February AO index of the previous winter ($R = 0.69$, $P < 0.01$). Bamzai [2003] showed that a lead-lag correlation between monthly snow cover and the AO index indicates that the winter AO is significantly correlated with snow cover in concurrent and subsequent spring months, mainly over Eurasia but also in North America. For the Arctic regions, Belchansky et al. [2004] showed that following high-index AO winters, spring melt tended to be earlier, although important and increasing unpredictability in snowmelt events was also observed.

3.3.8. Component 8: Labrador Peninsula

[47] Finally, component 8, representative of the Labrador Peninsula, shows no correlation with any teleconnection index.

3.4. Surface Pressure Connections to Explain the Role of Teleconnections With Respect to the Variability in the Date of Snow Cover Disappearance

[48] Finally, we have analyzed the relationship between the components of date of snow cover disappearance, the sea level pressures (SLP) and the temperatures. The purpose is to give a more direct knowledge about the role of the atmospheric circulation on the snow cover disappearance, also connecting with the physical mechanisms that relate both processes. Figures 7a and 7b shows the spatial distribution of correlations between the eight components and sea level pressure data during the months in which snow cover disappears for each area (indicated by A). The correlation between the average temperature and the sea surface pressure is also shown (indicated by B). To emphasize the processes responsible for the effect of teleconnections on the variability in the date of snow cover disappearance, the correlations between sea level pressures and those teleconnection indices significantly correlated with each component are also shown (indicated by C and D).

[49] Regarding component 1 (western Siberia), the spatial configuration of SLP indicates negative correlations with the April–May SLP in central Siberia and positive correlations with SLP in eastern Europe. This indicates that an earlier date of snow cover disappearance in these regions would be favored by dominant southern flows, determined by positive SLP in central Siberia and negative SLP in

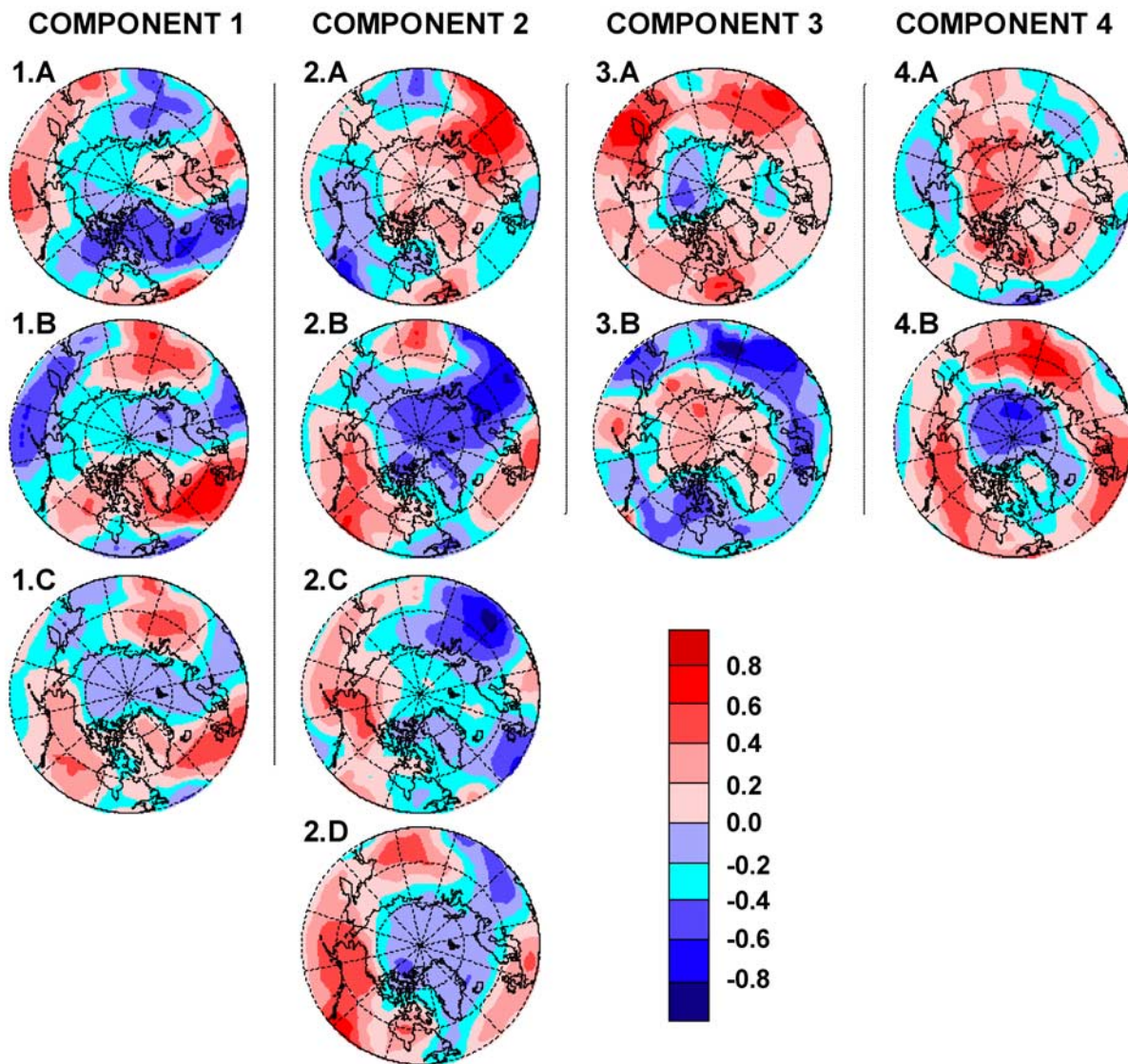


Figure 7a. (a) Spatial distribution of correlations between the North Hemisphere SLP and the components obtained from the date of snow cover disappearance (indicated by A), temperatures (indicated by B) and teleconnection indices (indicated by C and D). Component 1: 1.A, April–May SLP/component 1; 1.B, April–May SLP/April–May temperature in areas of component 1; 1.C, April–May SLP/AO (June of the previous year). Component 2: 2.A, April–May SLP/component 2; 2.B, April–May SLP/April–May temperature in areas of component 2; 2.C, April–May SLP/EA-WR (spring of the current year); 2.D, April–May SLP/SOI (June of the previous year). Component 3: 3.A, June SLP/component 3; 3.B, June SLP/June temperature in areas of component 3. Component 4: 4.A, April–May SLP/component 4, 4.B, April–May SLP/April–May temperature in areas of component 4.

eastern Europe (1.A). The correlation with temperature is of the opposite sign (1.B), higher temperatures being favored by southern flows. These spatial patterns of correlations with SLP are also clearly connected with the previous summer's AO index because the spatial distribution of correlations with the April–May SLP (1.C) resembles the SLP-temperature correlation pattern. Therefore negative (positive) June AO index of the previous year would produce north (south) flows in this region that significantly affect the air temperature and snow cover disappearance.

These results suggest the possibility of an early prediction of the date of snow cover disappearance in western Siberia.

[50] The correlation between component 2 (central Siberia) and the April–May SLP shows dominant negative values in the eastern areas of Siberia and positive values in western Siberia (2.A): an earlier/later snow cover disappearance is favored by southern flows in central Siberia, determined by positive SLP in central Siberia and negative SLP in western Siberia. The correlation pattern when considering temperatures is the opposite (2.B), there being

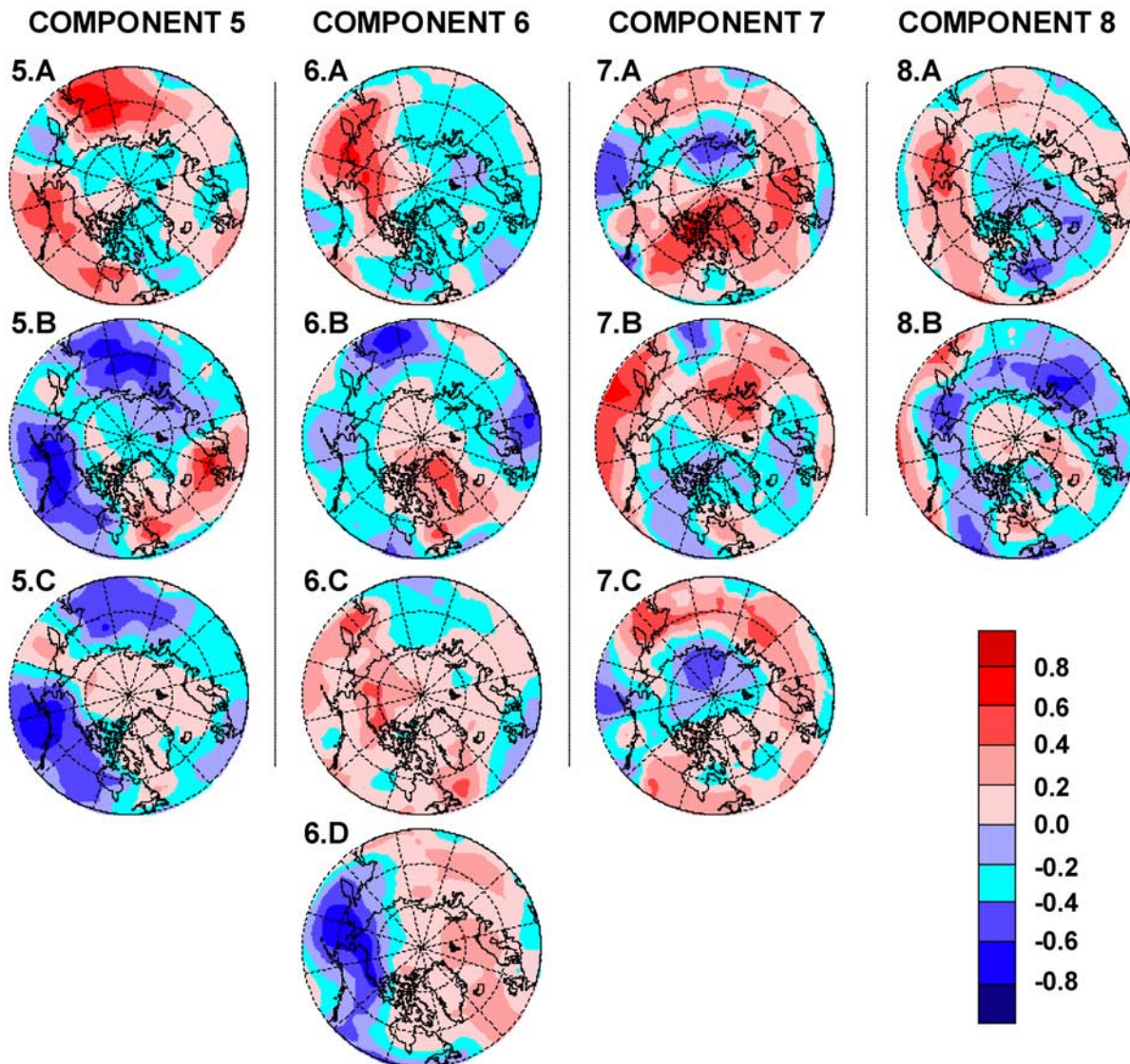


Figure 7b. Spatial distribution of correlations between the North Hemisphere SLP and the components obtained from the date of snow cover disappearance (indicated by A), temperatures (indicated by B) and teleconnection indices (indicated by C and D). Component 5: 5.A, May SLP/component 5; 5.B, May SLP/May temperature in areas of component 5; 5.C, May SLP/PNA (May of the current year). Component 6: 6.A, March–April–May SLP/component 6; 6.B, March–April–May SLP/March–April–May temperature in areas of component 6; 6.C, March–April–May SLP/SOI (spring of the current year); 6.D, March–April–May SLP/WP (spring of the previous year). Component 7: 7.A, June SLP/component 7; 7.B, June SLP/June temperature in areas of component 7; 7.C, June SLP/AO (winter of the current year). Component 8: 8.A, April–May SLP/component 8; 8.B, April–May SLP/April–May temperature in areas of component 8.

higher/lower temperatures as a consequence of south/north flows. Chart 2.C shows the correlation between the spring EA-WR pattern and the April–May SLP. There are strong negative correlations between the spring EA-WR pattern and SLP in large areas of central and western Siberia, indicating strong flows from the south (north) during positive (negative) EA-WR phases. This explains the negative correlation between the EA-WR and the variability in the date of snow cover disappearance in region 2, because

negative EA-WR values cause dominant north flows, which directly affects the temperatures of the region (low temperatures) and delays snow cover disappearance. A similar SLP configuration is recorded when considering the SOI of the previous summer, which noticeably affects the spring EA-WR and the dominant flows [Vicente-Serrano *et al.*, 2006a], thus explaining the negative correlation with component 2.

[51] For component 3, we found no correlation between snow cover disappearance and the teleconnection indices.

However, the correlation between the series of component 3 and SLP in June (the month in which snowmelt is usually recorded in the northern regions of Siberia) shows a clear pattern. High positive correlation values are found over large regions of central and western Siberia, and negative ones in the Arctic Ocean (3.A). The opposite pattern is observed when temperatures are analyzed (3.B). In this region a significant negative correlation between the date of snow cover disappearance and the temperature in June was found ($R = -0.67$, $p < 0.01$). This means that the snowmelt and temperature interannual variability is controlled by the intensity and direction of zonal flows rather than by the meridian flows characterizing most teleconnection patterns. The atmospheric pattern shown in 3.A indicates that higher (lower) temperatures and earlier (later) date of snow cover disappearance in region 3 are related to high (low) pressures in the northern Eurasian region, which cause east (west) flows in this area. Western flows give rise to warmer temperatures that may cause earlier snowmelt. The atmospheric configuration in 3.A and 3.B does not correspond to any general atmospheric circulation summer pattern, which may explain the negligible influence of the teleconnection indices on the variability in the date of snow cover disappearance in this region. Given the scarce data available for this remote region, not much is known about the influence of atmospheric circulation on the environmental processes in this area. *Vicente-Serrano et al.* [2006b] found that atmospheric circulation affects significantly the interannual variability of vegetation activity in this region, mainly by a summer pattern with a zonal configuration that is not related to any teleconnection index.

[52] For component 4, we also found no significant correlations with any teleconnection index, or with the average temperature in this area. The spatial complexity of component 4 makes it difficult to identify the variability in the date of snow cover disappearance/atmospheric circulation connections. Weak correlations and poorly defined spatial patterns were observed between component 4 and April–May SLP (4.A). In contrast, a clear zonal component is identified in the correlations between SLP and the surface temperature (4.B). In the region individuated by component 4 we found no relationship between temperatures and the date of snow cover disappearance. Therefore the physical connection of atmospheric circulation/temperature and the date of snow cover disappearance remain unclear in this area.

[53] For component 5 (southern Canada) there is a clear spatial pattern of correlations with the May SLP, with dominant positive correlations for the whole of Canada, mainly at latitudes of around 60°N (5.A). This atmospheric configuration explains the negative effect of the PNA on the date of snow cover disappearance in southern Canada. Positive values of the PNA cause negative SLP anomalies in western America (5.C), similar to the configuration of correlations between temperature and SLP (5.B). During negative phases of the PNA this spatial pattern of SLP causes dominant southern flows in most of North America, which may cause an earlier snowmelt related to southern flows and higher temperatures.

[54] The correlation between component 6 (northeastern Canada and the northern islands) and the March–April–May SLP values shows a very clear spatial pattern charac-

terized by strong positive correlations from the east of Eurasia to northern Canada (6.A). This indicates that high pressures in the Beringia region, which cause north flows, delay snowmelt processes in northeastern Canada. The correlation between temperature and SLP produces less well-defined spatial patterns (6.B), this being consistent with the fact that we found no relationship between average temperatures and the variability in the date of snow cover disappearance in these areas. The correlation between the spring SOI and SLP shows a similar pattern to that depicted in chart 6.A, although the correlation values are lower (6.C). The correlation between the spring WP and SLP also produces a very clear spatial pattern, with negative correlations between the WP and SLP from eastern Eurasia to northern Canada (6.D), which is opposite to that found for component 6. This explains the negative correlation found between the spring WP index and component 6: in fact, negative WP values are prone to cause positive pressures centered in Beringia, north flows and a later date of snow cover disappearance in areas represented by component 6.

[55] Component 7 (northern Canada) is positively correlated with the June SLP over a wide area from central Canada to Greenland (7.A). With high SLP values in this area, this spatial configuration would cause dominant north flows, which would affect region 7. This configuration might also cause a temperature decrease (7.B) and therefore delayed snowmelt. The positive correlation between component 7 and the winter AO could be due to a similar, although weaker, spatial pattern of SLP correlations. In any case, it is evident that region 7 is characterized by a well-defined spatial configuration of SLP that affects the variability in the date of snow cover disappearance. This spatial configuration, as in the case discussed for component 3, would not be well summarized by any teleconnection index.

[56] Finally, for region 8 (Labrador Peninsula) we found no correlation between the date of snow cover disappearance and the different teleconnection indices. However, there is a strong negative correlation with the April–May SLP in the north of the Labrador Peninsula (8.A). The opposite pattern is found between SLP and surface temperatures (8.B). This indicates that positive SLP in the north of the Labrador Peninsula would cause dominant southwest flows, which would cause higher temperatures and an earlier snowmelt in the Labrador Peninsula. In contrast, negative SLP in this area would cause dominant northern flows, colder temperatures and a later date of snow cover disappearance in this region. As before, this atmospheric influence was not recorded by means of the analyzed teleconnection indices, which suggests that in this area the teleconnection indices are not a good way of recording the atmospheric circulation patterns controlling the interannual variability in the date of snow cover disappearance.

4. Summary and Conclusions

[57] In this paper we have analyzed the spatial and temporal variability in the date of snow cover disappearance in the Northern Hemisphere between 1988 and 2003 and in connection with general atmospheric circulation. The principal patterns of variability in the date of snow cover disappearance were obtained from principal components analysis (PCA) in the S mode. This approach seems more

appropriate than considering average values within continental or regional boundaries since the spatial variability within the same geographical region is sometimes quite important (for example, we found four different and uncorrelated temporal patterns in Eurasia).

[58] We selected eight main regions where the temporal variability in the date of snow cover disappearance is homogeneous and analyzed its relationship to atmospheric circulation and temperatures. Temperature was found to have a direct impact on the variability in the date of snow cover disappearance in most areas. Moreover, the variability in the date of snow cover disappearance was found to be significantly related to atmospheric circulation variability, summarized by means of the teleconnection indices, over large areas: (1) to the AO in western Siberia, (2) to the EA/WR in central Siberia, (3) to the PNA in southern Canada, and (4) to the SOI and the WP in northwest America.

[59] More local atmospheric circulation patterns obtained from SLP were found to have a direct impact on the date of snow cover disappearance variability of other regions: (1) northern Siberia, (2) northernmost areas of Canada, and (3) Labrador Peninsula.

[60] We are aware that the correlation observed over a 16-year period may be irrelevant for some major modes of circulation, which exhibit regime shifts over a longer period. However, the results obtained are consistent with the findings of previous studies of snow cover variability for western Siberia [Banzai, 2003], central Siberia [Clark et al., 1999; Morinaga et al., 2003; Ueda et al., 2003] and North America [Gutzler and Rosen, 1992; Gutzler and Preston, 1997; Smith and O'Brien, 2001]. The analysis presented in this paper has also identified other areas where the variability in the date of snow cover disappearance is homogeneous. In Canada, some atmospheric circulation patterns (PNA and WP) affect the variability in the date of snow cover disappearance, and this is explained by their connection to different SLP that cause flows of different direction, which finally affect these regions during the snowmelt months (March, April and May). In northern Siberia, some areas of northern Canada and the Labrador Peninsula, no influence of the teleconnections has been found. Nevertheless, we did find that in these regions air temperature is mainly driven by SLP configurations characterized by zonal/meridian flows that are not related to any teleconnection index. The role of other minor teleconnection patterns, such as the Polar, the tropical/Northern Hemisphere and the Pacific Transition (Climate Prediction Center, <http://www.cpc.ncep.noaa.gov/data/teledoc/telecontents.shtml>), was also analyzed for these regions but no relationships were found. This would indicate that atmospheric circulation variability also plays an important role in explaining the date of snow cover disappearance in these regions, but that it is not summarized by the main teleconnection indices considered in this study.

[61] The global analysis reported here, and in particular the distinctions among the areas affected by the general atmospheric circulation and those controlled by temperature, may be very useful for snow modeling in hydrological models, GCMs and dynamic vegetation models based either on temperature only or on more complicated energy budget calculations requiring other input data such as albedo, long

wave radiation, atmospheric transmissivity, ground heat conduction and wind speed, etc.

[62] The results obtained in this paper may help in assessing carbon exchange in large boreal regions in which the gaseous flux interchanges between atmosphere and vegetation begin after snowmelt [Galen and Stanton, 1995; Ouerbauer et al., 1998]. Moreover, predicting the date of snow cover disappearance by means of teleconnection indices could be very useful for water resource management of Arctic rivers, particularly in regions such as western Siberia where a time lag between atmospheric circulation and snowmelt has been identified.

[63] **Acknowledgments.** The authors wish to acknowledge financial support from the projects Siberia II (5th Framework Program of the European Commission) and CGL2005-04508BOS (Ministerio de Educación y Ciencia, Spain, and FEDER).

References

- Aagaard, K., and E. C. Carmack (1989), The role of sea ice and other fresh water in the Arctic circulation, *J. Geophys. Res.*, *94*, 14,485–14,498.
- Albert, M., and G. Krajewski (1998), A fast, physically based point snowmelt model for use in distributed applications, *Hydrol. Processes*, *12*, 1809–1824.
- Ambaum, H. P., and B. J. Hoskins (2001), Arctic Oscillation or North Atlantic Oscillation?, *J. Clim.*, *14*, 3495–3507.
- Armstrong, R. L., and M. J. Brodzik (2001), Recent Northern Hemisphere snow extent: A comparison of data from visible and microwave sensors, *Geophys. Res. Lett.*, *28*, 3673–3676.
- Baldwin, M. P., and T. J. Dunkerton (1999), Propagation of the Arctic Oscillation from the stratosphere to the troposphere, *J. Geophys. Res.*, *104*, 30,937–30,946.
- Banzai, A. S. (2003), Relationship between snow cover variability and Arctic Oscillation index on a hierarchy of time scales, *Int. J. Climatol.*, *23*, 131–142.
- Barnett, L., U. Dümenil, E. 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.
- Barnston, A. G., and R. E. Livezey (1987), Classification, seasonality, and persistence of low-frequency atmospheric circulation patterns, *Mon. Weather Rev.*, *115*, 1083–1126.
- Barry, R. G., and A. M. Carleton (2001), *Synoptic and Dynamic Climatology*, Routledge, Boca Raton, Fla.
- Basnett, T. A., and D. E. Parker (1997), Development of the global mean sea level pressure data set GMSLP2, *Clim. Res. Tech. Note 79*, Hadley Cent., Meteorol. Off., Bracknell, U. K.
- Bednorz, E. (2002), Snow cover in western Poland and macro-scale circulation conditions, *Int. J. Climatol.*, *22*, 533–541.
- Belchansky, G. I., D. C. Douglas, and N. G. Platonov (2004), Duration of the Arctic sea ice melt season: Regional and interannual variability, 1979–2001, *J. Clim.*, *17*, 67–80.
- Briffa, K. R., P. D. Jones, and M. Hulme (1994), Summer moisture variability across Europe, 1892–1991: An analysis based on the Palmer drought severity index, *Int. J. Climatol.*, *14*, 475–506.
- Brown, R. D. (2000), Northern Hemisphere snow cover variability and change, 1915–1997, *J. Clim.*, *13*, 2339–2355.
- Brown, R. D., M. G. Hughes, and D. A. Robinson (1995), Characterizing the long-term variability of snow-cover extent over the interior of North America, *Ann. Glaciol.*, *21*, 45–50.
- Carrera, M. L., R. W. Higgins, and V. E. Kousky (2004), Downstream weather impacts associated with atmospheric blocking over the northeast Pacific, *J. Clim.*, *17*, 4823–4839.
- Cattell, R. B. (1966), The scree test for the number of factors, *Multivariate Behavioral Res.*, *1*, 245–276.
- Chapin, F. S., et al. (2000), Arctic and boreal ecosystems of western North America as components of the climate system, *Global Change Biol.*, *6*, 211–223.
- Clark, M. P., M. C. Serreze, and D. A. Robinson (1999), Atmospheric controls of Eurasian snow extent, *Int. J. Climatol.*, *19*, 27–40.
- Cohen, J., and D. Rind (1991), The effect of snow cover on climate, *J. Clim.*, *4*, 680–706.
- Corti, S., F. Molteni, and C. Brancovic (2000), Predictability of snow-depth anomalies over Eurasia and associated circulation patterns, *Q. J. R. Meteorol. Soc.*, *126*, 241–262.

- Díaz, H. F., and G. N. Kiladis (1992), Atmospheric teleconnections associated with the extreme phases of the Southern Oscillation, in *El Niño: Historical and Paleoclimatic Aspects of the Southern Oscillation*, edited by H. F. Díaz and V. Markgraf, pp. 7–28, Cambridge Univ. Press, New York.
- Dye, D. G. (2002), Variability and trends in the annual snow-cover cycle in Northern Hemisphere land areas, *Hydrol. Processes*, *16*, 3065–3077.
- Dye, D. G., and C. J. Tucker (2003), Seasonality and trends of snow-cover, vegetation index, and temperature in northern Eurasia, *Geophys. Res. Lett.*, *30*(7), 1405, doi:10.1029/2002GL016384.
- Essery, R., and P. Etchevers (2004), Parameter sensitivity in simulations of snowmelt, *J. Geophys. Res.*, *109*, D20111, doi:10.1029/2004JD005036.
- Esteban, P., P. D. Jones, J. Martín-Vide, and M. Mases (2005), Atmospheric circulation patterns related to heavy snowfall days in Andorra, Pyrenees, *Int. J. Climatol.*, *25*, 319–329.
- Fraedrich, K. (1994), An ENSO impact on Europe?, *Tellus, Ser. A*, *46*, 541–552.
- Frauenfeld, O. W., T. Zhang, R. G. Barry, and D. Gilichinsky (2004), Interdecadal changes in seasonal freeze and thaw depths in Russia, *J. Geophys. Res.*, *109*, D05101, doi:10.1029/2003JD004245.
- Fontaine, T. A., T. S. Cruickshank, J. G. Arnold, and R. H. Hotchkiss (2002), Development of a snowfall-snowmelt routine for mountainous terrain for the soil water assessment tool (SWAT), *J. Hydrol.*, *262*, 209–223.
- Foster, J. L., M. Owe, and A. Rango (1983), Snow cover and temperature relationships in North America and Eurasia, *J. Clim. Appl. Meteorol.*, *22*, 460–469.
- Fukutomi, Y., H. Igarashi, K. Masuda, and T. Yasunari (2003), Interannual variability of summer water balance components in three major river basins of northern Eurasia, *J. Hydrometeorol.*, *4*, 283–296.
- Galen, C., and M. L. Stanton (1995), Responses of snowbed plant species to changes in growing season length, *Ecology*, *76*, 1546–1557.
- Gershunov, A., and T. P. Barnett (1998), Interdecadal modulation of ENSO teleconnections, *Bull. Am. Meteorol. Soc.*, *79*, 2715–2725.
- Gong, G., D. Entekhabi, J. Cohen, and D. Robinson (2004), Sensitivity of atmospheric response to modeled snow anomaly characteristics, *J. Geophys. Res.*, *109*, D06107, doi:10.1029/2003JD004160.
- Goulden, M. L., et al. (1998), Sensitivity of boreal forest carbon balance to soil thaw, *Science*, *279*, 214–221.
- Grippa, M., N. Mognard, T. Le Toan, and E. G. Josberger (2004), Siberia snow depth climatology derived from SSM/I data using a combined dynamic and static algorithm, *Remote Sens. Environ.*, *93*, 30–41.
- Grippa, M., et al. (2005a), On the relationship between vegetation and snow indicators derived from remote sensing over central Siberia, *Geophys. Res. Lett.*, *32*, L21412, doi:10.1029/2005GL024286.
- Grippa, M., N. Mognard, and T. Le Toan (2005b), Comparison between the interannual variability of snow parameters derived from SSM/I and the Ob River discharge, *Remote Sens. Environ.*, *98*, 35–44.
- Groisman, P. Y., and D. R. Easterling (1994), Variability and trends of total precipitation and snowfall over the United States and Canada, *J. Clim.*, *7*, 1873–1891.
- Groisman, P. Y., T. R. Karl, and R. W. Knight (1994), Observed impact of snow cover on the heat balance and the rise of continental spring temperatures, *Science*, *263*, 198–200.
- Gutzler, D. S., and J. W. Preston (1997), Evidence for a relationship between spring snow cover in North America and summer rainfall in New Mexico, *Geophys. Res. Lett.*, *24*, 2207–2210.
- Gutzler, D. S., and R. D. Rosen (1992), Interannual variability of winter-time snow-cover across the Northern Hemisphere, *J. Clim.*, *5*, 1441–1447.
- Hair, J. F., R. E. Anderson, R. L. Tatham, and W. C. Black (1998), *Multivariate Data Analysis*, 799 pp., Prentice-Hall, Upper Saddle River, N. J.
- Häkkinen, S. (2000), Decadal air-sea interactions in the North Atlantic based on observations and modeling results, *J. Clim.*, *13*, 1195–1219.
- Halpert, M. S., and C. F. Ropelewski (1992), Surface temperature patterns associated with the Southern Oscillation, *J. Clim.*, *5*, 577–593.
- Heino, R., et al. (1999), Progress in the study of climatic extremes in northern and central Europe, *Clim. Change*, *42*, 151–181.
- Hock, R. (2003), Temperature index melt modelling in mountain areas, *J. Hydrol.*, *282*, 104–115.
- Hori, M. E., and T. Yasunari (2003), NAO impact towards the springtime snow disappearance in the western Eurasian continent, *Geophys. Res. Lett.*, *30*(19), 1977, doi:10.1029/2003GL018103.
- Hurrell, J. (1995), Decadal trends in North Atlantic Oscillation and relationship to regional temperature and precipitation, *Science*, *269*, 676–679.
- Hurrell, J., and H. Van Loon (1997), Decadal variations in climate associated with the North Atlantic Oscillation, *Clim. Change*, *36*, 301–326.
- Jolliffe, I. T. (1990), Principal component analysis: A beginner's guide. Part I: Introduction and application, *Weather*, *45*, 375–382.
- Jones, P. D., and A. Moberg (2003), Hemispheric and large-scale surface air temperature variations: An extensive revision and an update to 2001, *J. Clim.*, *16*, 206–223.
- Kaiser, H. E. (1958), The varimax criterion for analytic rotation in factor analysis, *Psychometrika*, *23*, 187–200.
- Karl, T. R., P. Y. Groisman, R. W. Knight, and R. R. Heim Jr. (1993), Recent variations of snow cover and snowfall in North America and their relation to precipitation and temperature variations, *J. Clim.*, *6*, 1327–1344.
- Kelly, R. E., and A. T. C. Chang (2003), Development of a passive microwave global snow depth retrieval algorithm for SSM/I and AMSR-E data, *Radio Sci.*, *38*(4), 8076, doi:10.1029/2002RS002648.
- Kerr, R. A. (2001), Getting a handle on the north's "El Niño," *Science*, *294*, 494–495.
- Kiladis, G. N., and H. F. Díaz (1989), Global climatic anomalies associated with extremes in the Southern Oscillation, *J. Clim.*, *2*, 1069–1090.
- Kitaev, L., A. Kislov, A. Krenke, V. Razuvaev, R. Martuganov, and L. Konstantinov (2002), The snow cover characteristics of northern Eurasia and their relationship to climatic parameters, *Boreal Environ. Res.*, *7*, 437–445.
- Krichak, S. O., and P. Alpert (2005), Decadal trends in the east Atlantic–west Russian pattern and Mediterranean precipitation, *Int. J. Climatol.*, *25*, 183–192.
- Kryjov, V. N. (2002), The influence of the winter Arctic Oscillation on the northern Russia spring temperature, *Int. J. Climatol.*, *22*, 779–785.
- Lamb, P. J., and R. A. Pepler (1987), North Atlantic Oscillation: Concept and application, *Bull. Am. Meteorol. Soc.*, *68*, 1218–1225.
- Lau, N. C. (1997), Interactions between global SST anomalies and the midlatitude atmospheric circulation, *Bull. Am. Meteorol. Soc.*, *78*, 21–33.
- Laxon, S., N. Peacock, and D. Smith (2003), High interannual variability of sea ice thickness in the Arctic region, *Nature*, *425*, 947–950.
- Livezey, R. E., and T. M. Smith (1999), Covariability of aspects of North American climate with global sea surface temperatures on interannual to interdecadal time scales, *J. Clim.*, *12*, 289–302.
- Lluch-Cota, D. B., W. S. Wooster, and S. R. Hare (2001), Sea surface temperature variability in coastal areas of the northeastern Pacific related to the El Niño–Southern Oscillation and the Pacific Decadal Oscillation, *Geophys. Res. Lett.*, *28*, 2029–2032.
- Mantua, N. J., S. R. Hare, Y. Zhang, J. M. Wallace, and R. C. Francis (1997), A Pacific interdecadal climate oscillation with impacts on salmon production, *Bull. Am. Meteorol. Soc.*, *78*, 1069–1079.
- Marshall, J., et al. (2001), North Atlantic climate variability: Phenomena, impacts and mechanism, *Int. J. Climatol.*, *21*, 1863–1898.
- McDowell, N. G., J. D. Marshall, T. D. Hooker, and R. Musselman (2000), Estimating CO₂ flux from snowpacks at three sites in the Rocky Mountains, *Tree Physiol.*, *20*, 745–753.
- Mitchell, T. D., T. R. Carter, P. D. Jones, M. Hulme, and M. New (2003), A comprehensive set of high-resolution grids of monthly climate for Europe and the globe: The observed record (1901–2000) and 16 scenarios (2001–2100), *Working Pap. 55*, Tyndall Cent. for Clim. Change Res., Norwich, U. K. (Available at http://www.tyndall.ac.uk/publications/working_papers/wp55.pdf)
- Mo, K. C., and R. E. Livezey (1986), Tropical-extratropical geopotential height teleconnections during the Northern Hemisphere winter, *Mon. Weather Rev.*, *114*, 2488–2515.
- Morinaga, Y., S. F. Tian, and M. Shinoda (2003), Winter snow anomaly and atmospheric circulation in Mongolia, *Int. J. Climatol.*, *23*, 1627–1636.
- Niebauer, H. J. (1998), Variability in Bering Sea ice cover as affected by a regime shift in the North Pacific in the period 1947–1996, *J. Geophys. Res.*, *103*, 27,717–27,737.
- Nijssen, B., G. O'Donnell, A. F. Hamlet, and D. P. Lettenmaier (2001), Hydrologic sensitivity of global rivers to climate change, *Clim. Change*, *50*, 143–175.
- 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.
- Ohmura, A. (2001), Physical basis for the temperature-based melt-index method, *J. Appl. Meteorol.*, *40*, 753–761.
- Ouerbauer, S. F., G. Starr, and E. W. Pop (1998), Effects of extended growing season and soil warming on carbon dioxide and methane exchange of tussock tundra in Alaska, *J. Geophys. Res.*, *103*, 29,075–29,082.
- Papineau, J. M. (2001), Wintertime temperature anomalies in Alaska correlated with ENSO and PDO, *Int. J. Climatol.*, *21*(13), 1577–1592.
- Peterson, B. J., R. M. Holmes, J. W. McClelland, C. J. Vorosmarty, R. B. Lammers, I. A. Shiklomanov, and S. Rahmstorf (2002), Increasing river discharge to the Arctic Ocean, *Science*, *298*, 2171–2173.
- Pulliainen, J., and M. Hallikainen (2001), Retrieval of regional snow water equivalent from space borne passive microwave observations, *Remote Sens. Environ.*, *75*, 76–85.

- Quinton, W. L., S. K. Carey, and N. T. Goeller (2004), Snowmelt runoff from northern alpine tundra hillslopes: Major processes and methods of simulation, *Hydrol. Earth Syst. Sci.*, *8*, 877–890.
- Rango, A., and J. Martinec (1995), Revisiting the degree-day method for snowmelt computations, *Water Resour. Bull.*, *31*, 657–669.
- Redmond, K. T., and R. W. Koch (1991), Surface climate and streamflow variability in the western United States and their relationship to large scale circulation indices, *Water Resour. Res.*, *27*, 2381–2399.
- Regonda, S. K., B. Rajagopalan, M. Clark, and W. Pitlick (2005), Seasonal cycle shifts in hydroclimatology over the western United States, *J. Clim.*, *18*, 372–384.
- Richman, M. B. (1986), Rotation of principal components, *J. Climatol.*, *6*, 29–35.
- Robinson, D. A., A. Frei, and M. C. Serreze (1995), Recent variations and regional relationships in Northern Hemisphere snow cover, *Ann. Glaciol.*, *21*, 71–76.
- Rogers, J. C., and H. Van Loon (1979), The seesaw in winter temperatures between Greenland and N. Europe. Part II: Some oceanic and atmospheric affects in middle and high latitudes, *Mon. Weather Rev.*, *107*, 509–519.
- Ropelewski, C. F., and M. S. Halpert (1987), Global and regional scale precipitation patterns associated with the El Niño/Southern Oscillation, *Mon. Weather Rev.*, *115*, 1606–1626.
- Ropelewski, C. F., and C. F. Jones (1987), An extension of the Tahiti–Darwin Southern Oscillation index, *Mon. Weather Rev.*, *115*, 2161–2165.
- 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., T. Yasunari, and J. Cohen (2004), Changes in the sub-decadal covariability between Northern Hemisphere snow cover and the general circulation of the atmosphere, *Int. J. Climatol.*, *24*, 33–44.
- Sankar-Rao, M., M. K. Lan, and S. Yang (1996), On the relationship between Eurasian snow cover and the Asian summer monsoon, *Int. J. Climatol.*, *16*, 605–616.
- 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.
- Schaefer, K., S. Denning, and O. Leonard (2004), The winter Arctic Oscillation and the timing of snowmelt in Europe, *Geophys. Res. Lett.*, *31*, L22205, doi:10.1029/2004GL021035.
- Schönwiese, C. D., and J. Rapp (1997), *Climate Trends Atlas of Europe Based on Observations, 1891–1990*, Springer, New York.
- Serrano, A., J. A. García, V. L. Mateos, M. L. Cencillo, and J. Garrido (1999), Monthly modes of variation of precipitation over the Iberian Peninsula, *J. Clim.*, *12*, 2894–2919.
- Serreze, M. C., J. E. Walsh, F. S. Chapin, T. Osterkamp, M. Dyurgerov, V. Romanovsky, W. C. Oechel, J. Morison, T. Zhang, and R. G. Barry (2000), Observational evidences of recent change in the northern high-latitude environment, *Clim. Change*, *46*, 159–207.
- Sheng, J., J. Derome, and M. Klasa (1998), The role of transient disturbances in the dynamics of the Pacific-North American pattern, *J. Clim.*, *11*, 523–536.
- Smith, N. V., and S. S. Saatchi (2004), Trends in northern latitude soil freeze and thaw cycles from 1988 to 2002, *J. Geophys. Res.*, *109*, D12101, doi:10.1029/2003JD004472.
- Smith, S. R., and J. J. O'Brien (2001), Regional snowfall distributions associated with ENSO: Implications for seasonal forecasting, *Bull. Am. Meteorol. Soc.*, *82*, 1179–1191.
- Stone, R. S. E., E. G. Dutton, J. M. Harris, and D. Longenecker (2002), Earlier spring snowmelt in northern Alaska as an indicator of climate change, *J. Geophys. Res.*, *107*(D10), 4089, doi:10.1029/2000JD000286.
- Thompson, D. W. J., and J. M. Wallace (1998), The Arctic Oscillation signature in the wintertime geopotential height and temperature fields, *Geophys. Res. Lett.*, *25*, 1297–1300.
- Thompson, D. W., M. P. Baldwin, and J. M. Wallace (2002), Stratospheric connection to Northern Hemisphere wintertime weather: Implications for prediction, *J. Clim.*, *15*, 1421–1428.
- Trenberth, K. E. (1990), Recent observed interdecadal climate changes in the Northern Hemisphere, *Bull. Am. Meteorol. Soc.*, *71*, 988–993.
- Trenberth, K. E. (1997), The definition of El Niño, *Bull. Am. Meteorol. Soc.*, *78*, 2771–2777.
- Trenberth, K. E., and J. W. Hurrell (1994), Decadal atmosphere-ocean variations in the Pacific, *Clim. Dyn.*, *9*, 303–319.
- Trenberth, K. E., and D. A. Paolino (1980), The Northern Hemisphere sea-level pressure data set: Trends, errors and discontinuities, *Mon. Weather Rev.*, *108*, 855–872.
- Ueda, H., M. Shinoda, and H. Kamahori (2003), Spring northward retreat of Eurasian snow cover relevant to seasonal and interannual variations of atmospheric circulation, *Int. J. Climatol.*, *23*, 615–629.
- Van Wijk, M. T., M. Williams, J. A. Laundre, and G. R. Shaver (2003), Interannual variability of plan phenology in tussock tundra: Modelling interactions of plan productivity, plant phenology, snowmelt and soil thaw, *Global Change Biol.*, *9*, 743–758.
- Vicente-Serrano, S. M., N. Delbart, T. Le Toan, and M. Grippa (2006a), El Niño–Southern Oscillation influences on the interannual variability of leaf appearance dates in central Siberia, *Geophys. Res. Lett.*, *33*, L03707, doi:10.1029/2005GL025000.
- Vicente-Serrano, S. M., M. Grippa, N. Delbart, T. Le Toan, and L. Kergoat (2006b), Influence of seasonal pressure patterns on temporal variability of vegetation activity in central Siberia, *Int. J. Climatol.*, *26*, 303–321.
- Wallace, J. M. (2000), North Atlantic Oscillation/annular mode: Two paradigms—One phenomenon, *Q. J. R. Meteorol. Soc.*, *126*, 791–805.
- Wallace, J. M., and D. S. Gutzler (1981), Teleconnections in the geopotential height field during the Northern Hemisphere winter, *Mon. Weather Rev.*, *109*, 784–812.
- Walter, M. T., E. R. Brooks, D. K. Cool, L. G. King, M. Molnau, and J. Boll (2005), Process-based snowmelt modeling: Does it require more input data than temperature-index modeling?, *J. Hydrol.*, *65*–75.
- Washington, R., A. Hodson, E. Isaksson, and O. MacDonald (2000), Northern Hemisphere teleconnection indices and the mass balance of Svalbard glaciers, *Int. J. Climatol.*, *20*, 473–487.
- White, D., H. Richman, and B. Yarnal (1991), Climate regionalization and rotation of principal components, *Int. J. Climatol.*, *11*, 1–25.
- Yang, D., D. Robinson, Y. Zhao, T. Estilow, and B. Ye (2003), Streamflow response to seasonal snowcover extent changes in large Siberian watersheds, *J. Geophys. Res.*, *108*(D18), 4578, doi:10.1029/2002JD003149.
- Yang, D., B. Ye, and D. L. Kane (2004), Streamflow changes over Siberian Yenisei river basin, *J. Hydrol.*, *296*, 59–80.
- 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, *J. Meteorol. Soc. Jpn.*, *69*, 473–489.
- Ye, B., D. Yang, and D. L. Kane (2003), Changes in Lena River streamflow hydrology: Human impacts versus natural variations, *Water Resour. Res.*, *39*(7), 1200, doi:10.1029/2003WR001991.
- Ye, H. (2000), Decadal variability of Russian winter snow accumulation and its associations with Atlantic sea surface temperature anomalies, *Int. J. Climatol.*, *20*, 1709–1728.
- Ye, H. (2001a), Characteristics of winter precipitation variation over northern central Eurasia and their connections to sea surface temperatures over the Atlantic and Pacific oceans, *J. Clim.*, *14*, 3140–3155.
- Ye, H. (2001b), Quasi-biennial and quasi-decadal variations in snow accumulation over northern Eurasia and their connections to the Atlantic and Pacific oceans, *J. Clim.*, *4*, 4573–4584.
- Ye, H., and J. R. Mather (1997), Polar snow cover changes and global warming, *Int. J. Climatol.*, *17*, 155–162.
- Ye, H., H. Cho, and P. E. Gustafson (1998), The changes in Russian winter snow accumulation during 1936–1983 and its spatial patterns, *J. Clim.*, *11*, 856–863.
- Zhou, L. M., C. J. Tucker, R. K. Kaufmann, D. Slayback, N. V. Shabanov, and R. B. Myneni (2001), Variations in northern vegetation activity inferred from satellite data of vegetation index during 1981 to 1999, *J. Geophys. Res.*, *106*, 20,069–20,083.
- Zorita, E., V. Kharin, and H. Von Storch (1992), The atmospheric circulation and sea surface temperature in the North Atlantic area in winter: Their interaction and relevance for Iberian precipitation, *J. Clim.*, *5*, 1097–1108.

M. Grippa and T. Le Toan, CESBIO, 18 avenue Edouard Belin, BPI 2801, F-31401 Toulouse, France.

N. Mognard, Laboratoire d'Etudes en Géophysique et Océanographie Spatiales, 18 avenue Edouard Belin, F-31401 Toulouse, France.

S. M. Vicente-Serrano, Instituto Pirenaico de Ecología, CSIC, Campus de Aula Dei, P.O. Box 202, E-50080 Zaragoza, Spain. (svicen@ipe.csic.es)