Paper V

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The influence of snow conditions on wintertime extratropical atmospheric variability – a study based on numerical simulations

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ABSTRACT

Two ensemble simulations have been performed with a global climate model of the atmosphere for the period 1972-1997 to quantify the impact of realistic snow conditions on the large scale variability of the atmosphere. The same prescribed sea surface temperature (SST) and sea ice have been used in both ensembles. Snow conditions, on the other hand, have been prescribed in one ensemble, but modeled in the other. The seasonal long-term means of atmospheric variables, like sea level pressure, differ negligibly between the two ensembles, while different snow cover representation gives large differences in the atmospheric variability over Eurasia and in the regions associated with the North Atlantic Oscillation (NAO). While anomalies in SST influence quasi-decadal NAO variability, it is found that the variability of the NAO on shorter time scales (2-4.5 years) is clearly modulated by realistic interannual snow forcing, mainly through variation in the Azores High. On seasonal time scale it is found that realistic variability of Eurasian snow cover can contribute to simultaneous changes of atmospheric thermal states in the same area at the surface and in lower troposphere. A significant negative correlation is found between wintertime thermal variation over the Azores High and Eurasian snow cover in autumn, but no similar correlation is found for the Icelandic Low. This indicates that snow variation over Eurasia can influence the wintertime NAO through modulation of the Azores High on a seasonal time scale. Extensive Eurasian snow cover in autumn heralds the negative wintertime NAO, and vice versa. The internal nonlinear processes in the atmosphere, causing the modulation effects on the NAO by snow forcing, is so far unclear. Hence, the nonlinear atmospheric response to snow forcing needs further theoretical considerations, such as from the point of the view of wave-wave interaction, wave-mean flow interaction and dispersion/propagation of wave fluxes.

1. Introduction

During recent years considerable efforts have been put on studying the impact of snow anomalies in early seasons on subsequent wintertime atmospheric variability, especially over extratropical North Atlantic. The Arctic Oscillation (AO) is an annular mode in the atmosphere, this primary mode predominates the extratropical Northern Hemisphere from the surface to the lower stratosphere (Thompson and Wallace 1998). The North Atlantic Oscillation (NAO), which is an important teleconnection pattern between the Icelandic Low and the Azores High, shows a regional surface signal of the AO pattern (Hurrell 1995; Baldwin and Dunkerton 1999). The AO/NAO is most dominant and recurrent in winter. Strong evidence is provided by atmospheric general circulation models (AGCMs) that the variation of the AO/NAO arises from the atmospheric nonlinear processes (Osborn et al. 1999; Paeth et al. 2003). However, forcing from underlying surface and the stratosphere can modify and amplify the variability of the AO/NAO (Walter and Graf 2002; Hoerling et al. 2001 and 2004; Baldwin et al. 1999; Ambaum and Hoskins 2002). In this paper, the influence of snow cover on the NAO is investigated.

Seasonal lagged connection between Eurasian snow cover and the NAO has been studied by Cohen and Entekhabi (1999). They pointed out that the cooling effect of Eurasian snow cover in autumn affects variation of wintertime NAO mainly through the anomalies in the strength and location of the Siberian high. Such lagged linkage appears significantly for the anomalies of Eurasian snow cover in October (Saito and Cohen, 2003). From dynamical investigations of upward wave activity fluxes (WAF), it has been indicated that enhanced vertical wave propagation takes place prior to the formation of the wintertime NAO/AO (Karoda and Kodera 1999). On the basis of WAF propagation under different snow conditions in early season (Cohen et al. 2002), a downward shift of zonal-mean zonal wind is expected in lower troposphere and at the surface in the subsequent winter through tropospheric-stratospheric coupling (Baldwin and Dunkerton 1999; Saito et al. 2001; Gong et al. 2003). In this way the autumn Eurasian snow cover can influence the variation of wintertime NAO. On the other hand, Kumar and Yang (2003) reported that the influence of snow on the variability of the extratropical circulation was confined to the lower troposphere through positive snow albedo feedback. Clearly, further studies are needed in order to understand and explain the impacts of snow variations on the extratropical atmospheric variability.

It is also documented that the observed NAO index contains significant variations at quasi-biennial and quasi-decadal periods (Hurrel et al. 2003; Saito and Cohen 2003). However, there are fewer investigations on the linkage between Eurasian snow conditions and wintertime NAO on these time scales than that on seasonal time scale.

Snow is an internal variable of the climate system. Variations of snow cover and snow depth are largely determined by anomalies of atmospheric circulation. For instance, Kodera et al. (1999) pointed out that the decreasing trend of snow over Siberia, associated with local warming, is strongly correlated with changes in the AO/NAO index. However, it is observed that snow cover could be seasonally persistent and in this way exert a strong forcing anomaly on the atmospheric circulation through the exchange of heat at the surface, diabatic heating and the hydrologic cycle. For example, Yeh et al. (1983) found that the removal of snow cover led a reduction of surface albedo, resulting in increased absorption of solar short-wave radiation, thus heating the ground surface and allowing for more evaporation to occur. It has been earlier pointed out that a strong simultaneous correlation exists between wintertime snow cover and surface temperature over North America, while wintertime surface temperature over Eurasia is significantly correlated with local autumn snow cover (Foster et al. 1983). Several statistical studies have documented an inverse relation between the strength of the Indian summer monsoon and preceding Eurasian snow cover and depth, especially snow anomaly in western Eurasia (Hahn and Shukla 1976; Kripalani et al. 1996). Barnett et al. (1989) explained that a positive Eurasian snow cover anomaly in winter/spring could weaken the summer monsoon circulation basically because of reduced land-sea temperature contrast caused by a preceding heavy snow anomaly.

Due to the importance of snow in the climate system, it is well known that proper representation of snow cover and snow depth variations in AGCMs is vital for climate simulations. The climatological distribution of snow cover and depth are represented fairly well by models (Foster et al. 1996), while it is also found that there are some difficulties for models to simulate well snow cover and snow depth during transition seasons of the spring and fall. Watanabe and Nitta (1998) studied the extreme event of sparse Eurasian snow in winter 1988/1989 and found that the Eurasian snow changed little between the simulations with prescribed and climatological SST forcings respectively. This implies that there is a common problem for AGCMs to represent realistic snow variation. From model simulations Schlosser and Mocko (2003) indicated that improved representation of snow conditions, by using assimilation of observed snow depth data during model integration, skillfully can capture the spatial pattern and temporal variations of near-surface air temperature at both local and remote scales. However, at present there are few studies that have estimated the temporal variability of model snow cover and snow depth in model simulations against observations and the effect of such variations on atmospheric variability. In this study, we try to perform AGCM simulations by using observed and modeled variability of snow cover in the model.

By using statistical analysis and numerical simulations, an attempt is made to document how the observed interannual snow variations modulate the atmospheric variability in boreal winter on different timescales. The article is organized as follows. The observational data, the atmospheric general circulation model, snow forcing and the design of model simulations are introduced in Section 2. Section 3 presents the climatological properties of snow conditions and simulated atmospheric circulation parameters to illustrate the ability of the model. The geographical pattern and temporal variability of the NAO is described in Section 4, devoting attention to the frequency modulation of the NAO variability forced by snow cover. The seasonal response on atmospheric thermal states from the interannual snow forcing is described in Section 5, and a summary and discussion are given in Section 6.

2. AGCM simulations and observational data

2.1 Model and snow parameterization

The AGCM used in this study is the third version of ARPEGE/IFS Climate model (Déqué et al. 1994, 1998). It is a spectral model with horizontal triangular truncation (Hortal 1998). In our experiments we have used the model with 63 waves (T63) and 31 vertical levels. Because the technique of reduced Gaussian grids is used in the model (Hortal and Simmons 1991), the horizontal grid spacing is approximately 2.8° in latitude and longitude.

Treatment of snow mass and associated snow fraction and albedo is employed in the ISBA (interaction among soil, biosphere and atmosphere) scheme (Douville et al. 1995a, b). The mean snow water equivalent (SWE) is governed by a prognostic equation which traditionally has been used in most AGCMs. Snow albedo exhibits an exponentially decrease in time if the snowmelt rate is positive, otherwise, a weak linear decrease is imposed during cold days. A snowfall refreshes the albedo back to maximum value when it exceeds a threshold value of SWE. The total snow cover fraction is closely related to SWE and takes account of the effect of vegetation and surface roughness. Snow density is assumed to be constant with depth and to decrease exponentially with time from maximum to minimum values. After a new snowfall, the snow density is recalculated as the weighted average of the previous density and that of new snow. The diagnostic variable of snow depth is determined from SWE and snow density. The effect of a snow reservoir is considered over Greenland, Himalaya, Rocky mountains and Antarctia, where the persistent ice cover over land is represented by a large snow depth and SWE.

Hence, in the present snow parameterization scheme, the variable SWE links all the state variations of snow. This gives us a hint on how to design our experiments.

2.2 Model simulations

In order to reproduce climate variability during last several decades and to explore the large-scale atmospheric response to improved snow conditions, we have generated two ensemble simulations (Table 1). The first ensemble of simulations is a control ensemble named CTL, in which boundary conditions (e.g. vegetation and its roughness length, deep soil temperatures, land ice extent and surface emissivity) are set to climatological values everywhere over land. Over sea observed interannual variation for SST and sea ice is used, and prescribed monthly global SST fields from 1972 to 1997 are employed as the major source of boundary forcing. The SST data is provided by a blend of two reconstructed datasets from GISST (Global sea Ice and Sea Surface Temperature) from 1950 to 1982 and from 1983 to 1997 respectively (Reynolds and Smith, 1994). Moreover, the latest dataset is constructed from both in situ and satellite observations using an optimum interpolation technique. Snow amount evolves freely in the model integrations of the ensemble CTL.

In the second ensemble, named SNS, the integrations are run with the same global distribution of observed SST and sea ice, but the representation of snow amount is different. It is here supposed that a simple positive proportional relationship exists between monthly climatological snow mass and observed snow cover fraction at every grid point over land, except for grid points with perpetual ice cover, i.e. the larger snow cover fraction, the more snow mass. A goal of this paper lies in evaluation of the atmospheric response in the simulations to more realistic snow conditions, rather than rigorously emphasizing the correct observed snow mass, which is not available at all. Therefore, in order to keep consistency and to decrease the chance of changing the model energy, the climatological snow amount, used in the mentioned proportional relationship, is derived from the control ensemble. The monthly snow mass in northern Hemisphere is then constructed from the observed snow cover by using the proportional relation. One advantage for doing this is no consideration about the variation of model snow density due to precipitation, snowmelt and sublimation when snow mass is constructed. In other studies the snow density is assumed to be constant at each grid point, and the snow amount is calculated according to constant snow density and variable snow depth (e.g. Watanabe and Nitta 1998; Schlosser and Mocko 2003). Another advantage from using the proportional relation is that it is possible for the model to keep the information on the temporal variability of observed snow cover during the integrations. The daily snow forcing, by temporal interpolation in SNS, is put into the model every five days after 1972, and the snow is allowed to evolve freely in the ISBA scheme for the subsequent time periods until new forcing is inserted. The characteristics of the new snow conditions are explored in the next section.

Each ensemble consists of 5 independent realization members with slightly different initial conditions. The integration period is 1972-1997. Outputs of ensemble means spanning the whole period are considered.

GCM simulation	SST specification	Snow specification	numbers	period
CTL	observed	free	5	1972-1997
SNS	observed	fixed	5	1972-1997

Table 1: Design of the two simulation ensembles CTL and SNS

2.3 Observational data

To identify realistic snow conditions, the weekly snow cover data from 1972 to 2000 from the National Snow and Ice Data Center (NSIDC) has been used (Armstrong and Brodzik 2002). The climatology and variations of the atmospheric states simulated by the model are compared with the corresponding observational data obtained from European Centre for Medium-Range Weather Forecasts (ECMWF) (Uppala et al. 2005; ECMWF 2002). The observed thermodynamic fluxes are provided by NCEP/NCAR reanalysis data (Kalnay et al. 1996). The Canadian Online Snow Atlas is used to test simulated snow depth (Brown et al. 2003). The wintertime NAO pattern is defined as the leading mode of sea level pressure (SLP) over December, January and February (DJF) and the NAO index used is the principal component of the leading mode (Hurrell, 1995).

3. Climatology

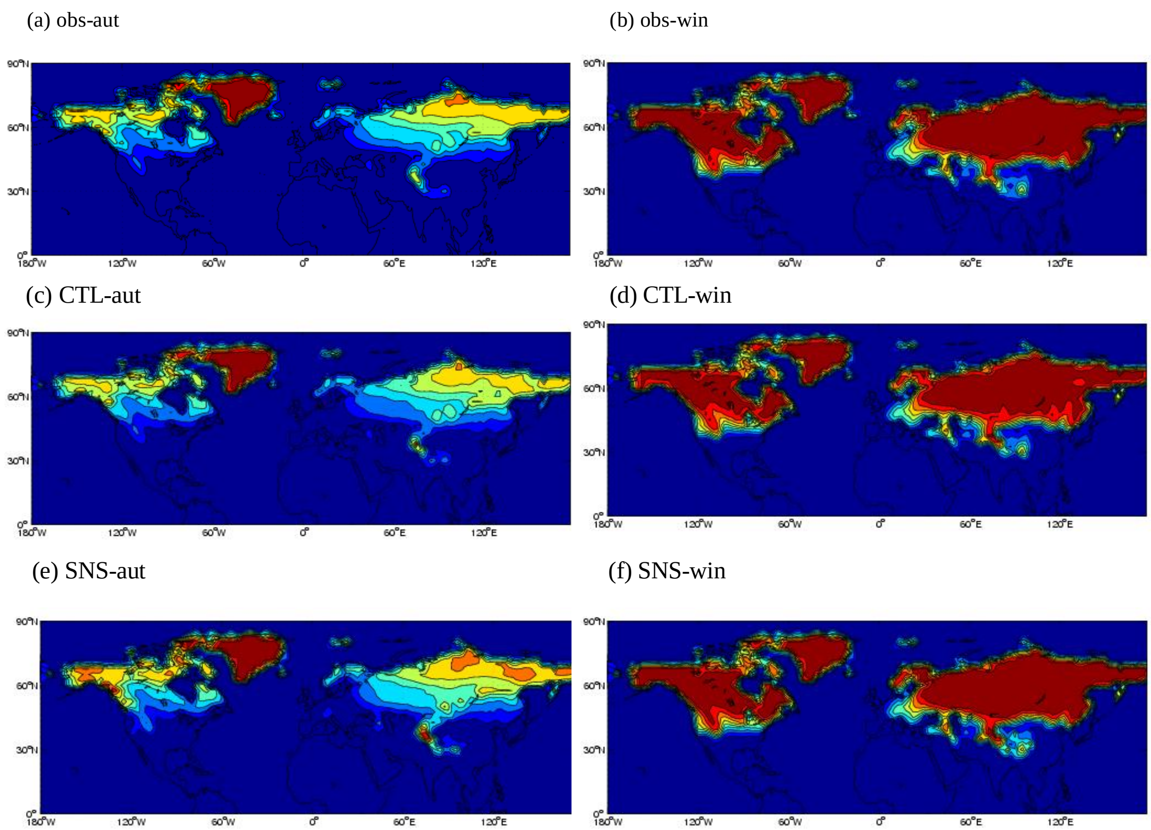

Figure 1. Long-term mean snow cover fraction in autumn (left panel) and winter (right panel) for observation and each set of ensemble (CTL and SNS) respectively.

To demonstrate the model's snow characteristics, the spatial distribution of long-term mean snow cover fraction in cold seasons is first presented (Fig. 1). The spatial distribution of snow cover in both ensembles is overall in good agreement with observations, except that the snow cover in CTL is obviously underestimated over Eurasia in autumn (Fig. 1b), while it is a little overestimated in SNS simulation over Siberia (Fig. 1c). In winter the snow cover exhibits similar simulated differences, but they shift southward (Fig. 1e and f). This consistent underestimation of snow mass in CTL is determined by the AGCM itself (Foster et al. 1996). The large snow cover over Himalaya and Rocky Mountains in the model is the regional effect of snow accumulation in high mountains.

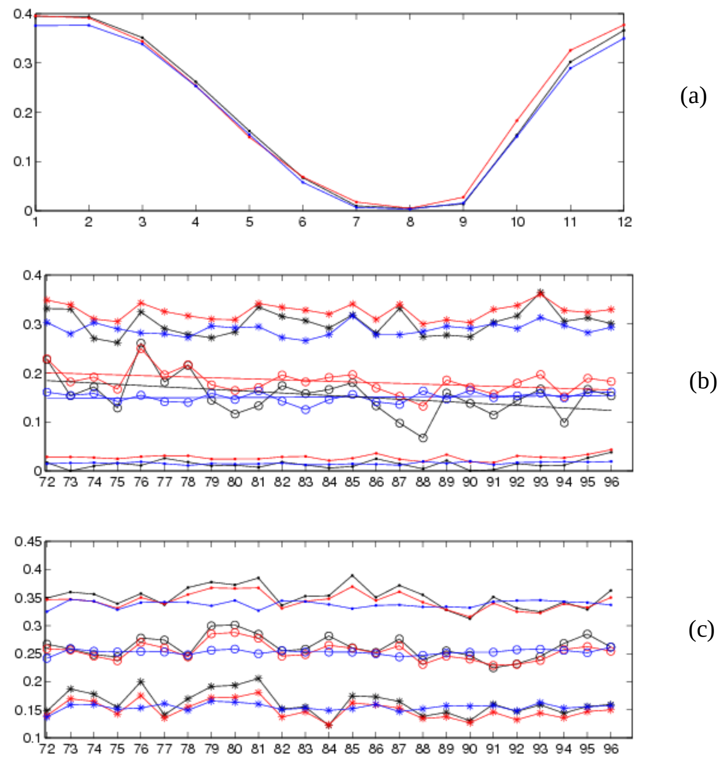

Figure 2. Long-term mean monthly variations of area-averaged snow cover over Eurasia (20°N-75 °N, 0-177.5 °E) (a) and interannual variations in autumn (b) and spring (c). Black line with marks represents the observed variation, blue and red lines are CTL and SNS simulations respectively, the lines with point, circle and star reflects the spatial snow extend or retreat in different months. The lines without marks in (b) are the tendency of snow variations in October.

In order to clearly see the temporal snow variation in the Eurasian continent, the area-averaged monthly snow cover over Eurasia (20°N-75°N, 0-177.5°E) is calculated. In both ensembles, the monthly area-averaged seasonal snow cover over Eurasia resembles the observed means (Fig. 2a), but the snow cover in SNS is systematically more than observed in autumn and slightly less in spring. For CTL there is evidently less snow in winter than observed. The spatial distribution and temporal variation of the snow cover seem to be well simulated in both ensembles (Fig. 1). However, the interannual variability of the snow cover differs substantially between CTL and SNS. The interannual variation of snow cover in SNS is similar to reality (Fig. 2b and c). For example, the temporal correlation of snow cover in October between SNS and observation is 0.9953. The decreasing trend in the snow cover in October is also in agreement with observations. This proves that the proportional relation to construct snow amount really keep the information on the interannual variation of observed snow cover in SNS ensemble. The interannual variability of snow in CTL is on the other hand underestimated.

Snow cover and snow depth variations are inherently related. Due to the lack of observed snow depth data, we simply compare the model climatological snow depth in SNS with the product from Canadian Online Snow Atlas (Fig. 3). The distribution of climatological snow depth in SNS agrees well with the observations both in January and July. This means that the seasonal cycle of snow depth is well performed in SNS. It is earlier found that the simulated snow depth is nearly always underestimated over mountains and forest regions (Foster et al. 1996), while the snow depth in parts of the Rocky Mountains is considerably larger in this model than in observations. Additional comparative analyses of simulated snow depth might be found in Douville et al. (1995).

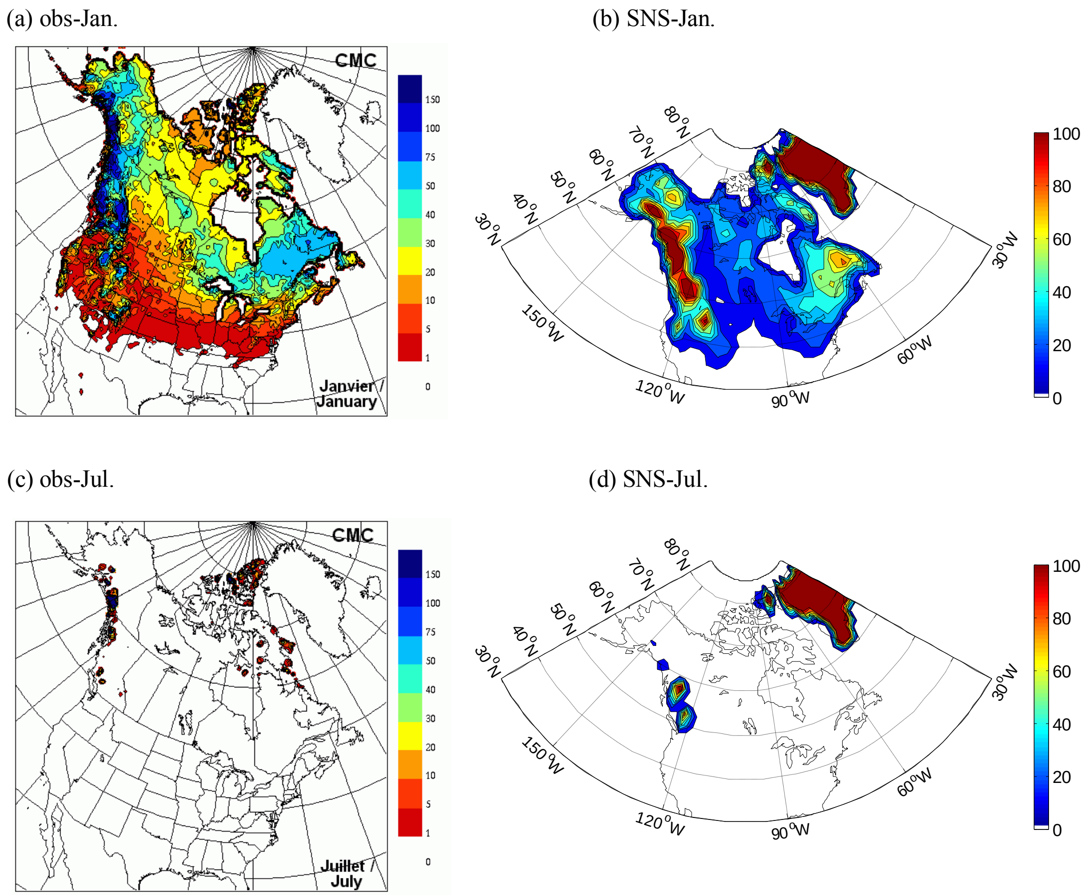

Figure 3. Long-term mean snow depth over North America in Jan. (a and c) and July (b and d). Left panel is got from Canadian Online Snow Atlas (http://www.socc.ca/nsisw/atlas/index.html), right panel is from SNS simulation (30°N-85°N, 170°W --30°W). The color interval is 10cm on right panel, unit: cm.

From the presentation of the climatological distributions of snow and the temporal variations of snow cover, it can be seen that the snow mass inserted into the model is reasonable. Therefore the climatological mean of atmospheric variables should not differ much between the two ensembles. The slight systematic biases of snow forcing and the similar interannual variability of the snow cover to reality in SNS may, however, affect the result of simulations. Hence, the influence of snow cover on the atmospheric variability may differ a lot between the two ensembles. Let us take DJF SLP fields as an example to illustrate this point.

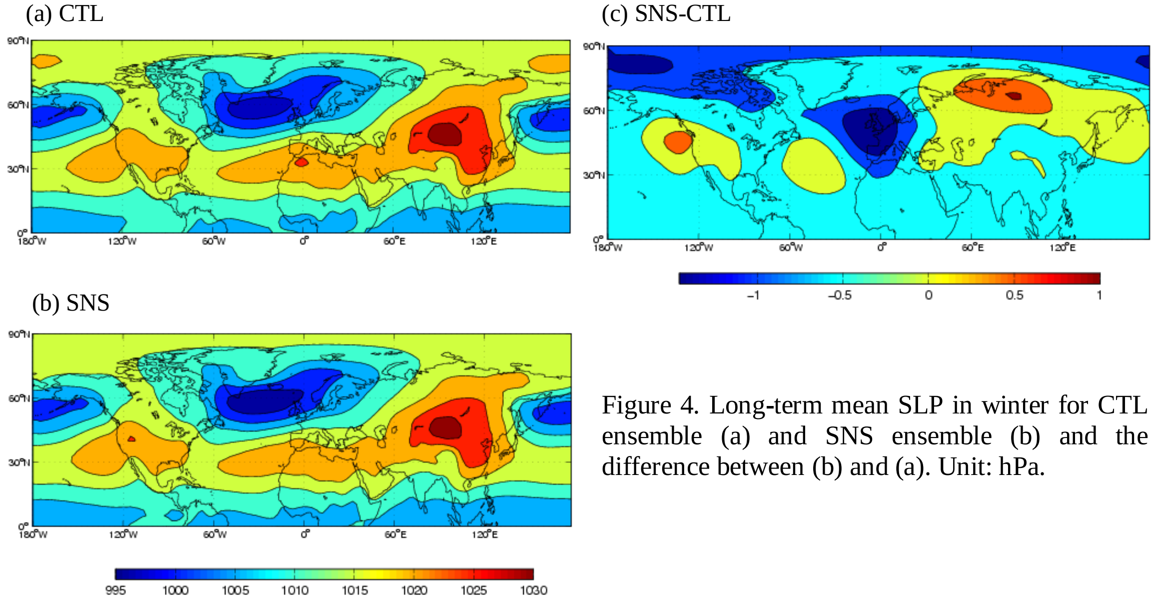

Figure 4. Long-term mean SLP in winter for CTL ensemble (a) and SNS ensemble (b) and the difference between (b) and (a). Unit: hPa.

It is clearly seen from Fig. 4a and 4b that the DJF distribution of climatological SLP mean is almost identical for the two ensembles, and also well in agreement with observation (not shown). The maximum difference between the two ensembles is up to 1.5 hPa and locates mainly over Eurasia and North Atlantic (Fig. 4c). Accordingly, the difference in long-term mean between the two ensembles is nearly negligible. However, the standard deviation of the DJF SLP differs a lot between the two ensembles. Comparing with atmospheric variability in CTL (Fig. 5a), interannual snow variation in SNS strengthens atmospheric variability between Ural Mountains and the Iranian plateau and weakens atmospheric variability over the North Atlantic sector (Fig. 5b).

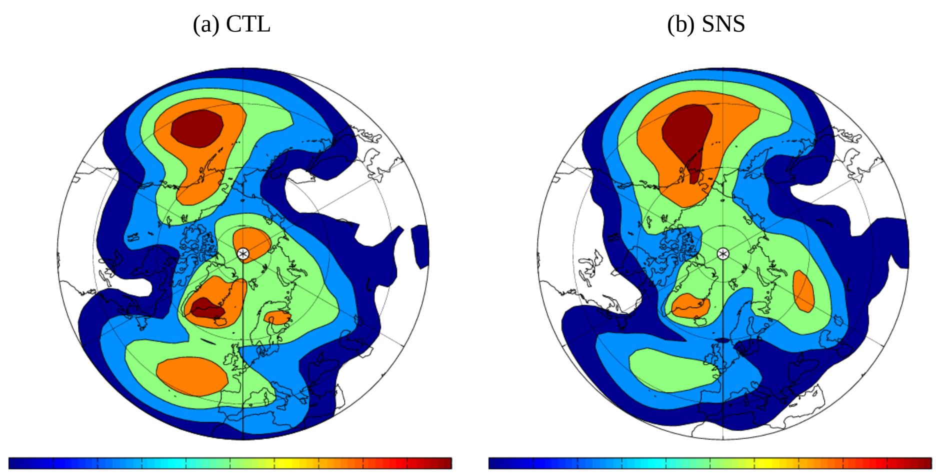

Figure 5. Standard deviation of the DJF SLP for the ensembles of CTL (a) and SNS (b).

Therefore, the characteristics of snow and atmospheric variables in our simulations indicate that the interannual fluctuation of snow cover can influence the spatial variability of the NAO. Cassou and Terray (2001) have documented the good ability for ARPEGE model to simulate the wintertime low-frequency atmospheric variability over the North Atlantic European sector. Hence, in the following the detailed impact of snow cover on the NAO is studied further.

4. Reproduction and frequency modulation of the NAO

Studies indicate that the NAO pattern is governed by nonlinear internal processes in the atmosphere. This might be seen through the fact that the NAO pattern is still present in AGCMs simulations forced by seasonal varying climatological SST (see Introduction). It is thus an inherent oscillation in the atmosphere, which is also proved in our simulations. The spatial pattern and amplitude of the NAO are typically well simulated in our AGCM experiments forced with prescribed SST and snow (Fig.6b and 6c), and explains more than 50% of the total variance, comparable to what is observed. The variance contribution here seems higher than that in Hurrel et al. (2003), this is mainly due to the different period investigated and possible more contribution of the NAO to extratropical atmospheric variability in last thirty year than that in the first half of last century. Therefore, in our study the basic dominant mode exists both with or without evident interannual snow variation.

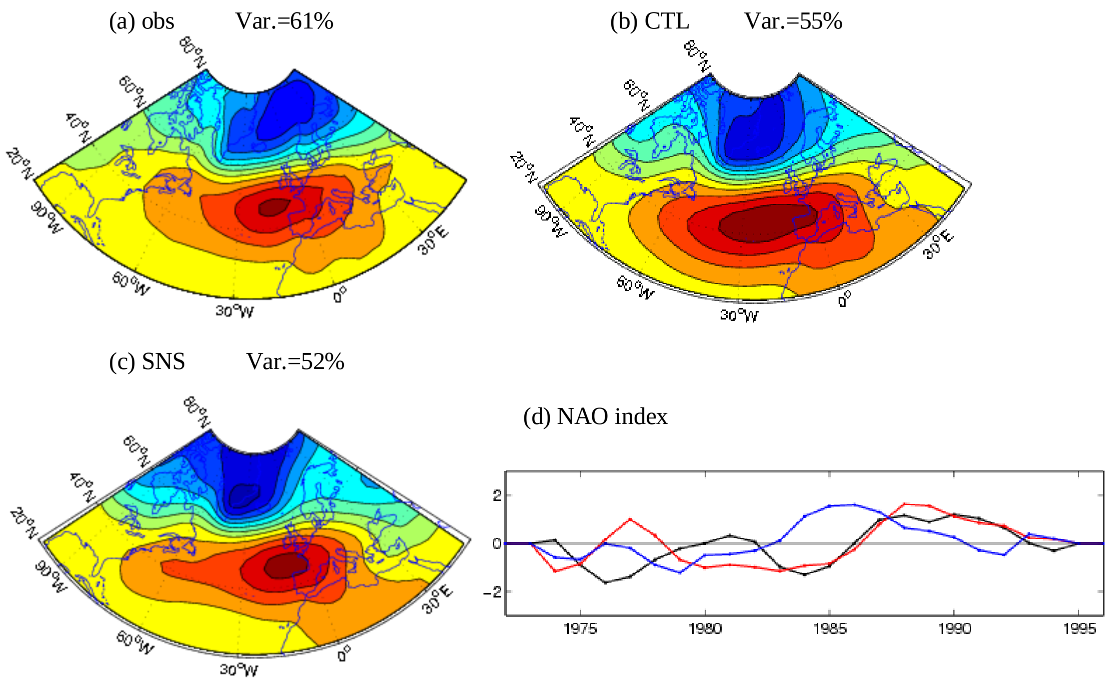

Figure 6: NAO pattern for observation (a), CTL (b) and SNS (c) simulations. (d) Standardized DJF NAO index for observation (black line), CTL (blue line) and SNS ensemble (red line) with low-pass filtering of 4.5 year. The indicated year corresponds to the Dec. of the winter season. The correlate coefficient of the NAO between SNS and ERA is 0.5 at a 98% confident level, insignificant 0.05 between CTL and ERA.

On the other hand, the temporal variability of the NAO is very different in the two ensemble simulations. The temporal variability of the NAO in CTL with 4.5-year low-pass filtering shows little coherence with reality (Fig. 6d), while the NAO index in SNS shows a correlation of 0.5 with observations at a 98% confidence level. Such an improved consistency of the NAO index implies that the interannual snow variation can influence the variability of the NAO.

In order to study the modulation effects of the NAO index with snow forcing, the power spectral estimation smoothed with a Hanning window is used next. The results in the following have passed the significant test at 95% confidence level. The observed NAO index contains a broad spectrum of variations with significant variance at quasi-biennial (2.5-3.5 years) and quasi-decadal (6.25-year) periods (Fig. 7a) (see Introduction). In CTL there are two main periods of the NAO on shorter time scale, around 2 and 4 years (Fig. 7b). They are modulated to be extended in SNS (Fig. 7c) except for the period of 3 years, which will be mentioned later. The longer main period of 8.3 years is similar in SNS and CTL, although the interannual snow variation (SNS) slightly strengthens this very low frequency oscillation. Hence, it seems that the large frequency modulation of snow forcing on the simulated NAO is focused on the time scale of 2-4.5 years.

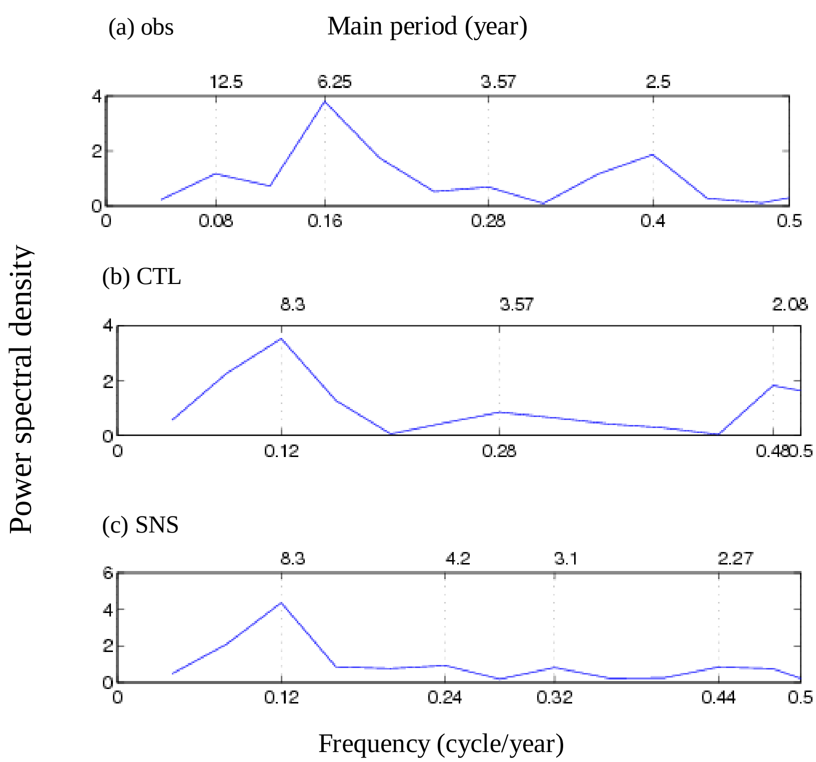

Figure 7. Power spectrum estimate on the DJF NAO index without filtering for observation (a), for CTL (b) and SNS (c) ensembles in the1972-1997 period.

Paeth et al. (2003) have pointed out that the observed interannual NAO fluctuations are barely influenced by the global SST, while the variability of the NAO is well reproduced by the SST on the decadal time scales. Hurrel et al. (2003) have also mentioned that the global SST can be expected to modulate NAO variability on longer time scales. Due to the shorter time series investigated in this study, our results associated with external forcings imply that global SST seems to modulate NAO variability at the quasi-decadal period.

A question arises here: how does interannual variation of snow forcing modulate the variability of the NAO? The power spectrum analysis is performed again on the DJF area-averaged anomaly of SLP over a North Atlantic area covering the normal position for the Azores High (55°W-0°E, 27.5°N-47.5°N) and an area in the position of the Icelandic Low (40°W-0°E, 55°N-70°N) respectively to seek the answer.

Fig. 8 presents the power spectral estimation over the Azores High. By comparing Fig. 8a with Fig. 7a, it can be clearly seen that the main periods of the observed time series agree well with that of the observed NAO index with respect to quasi-biennial to quasi-decadal periods. The energy densities are at the same level as well. Due to significant high power spectral density at the main periods of 2-4 years (Fig. 8a), it seems that the variability of the Azores High is predominant on shorter time scales associated with the variability of the NAO.

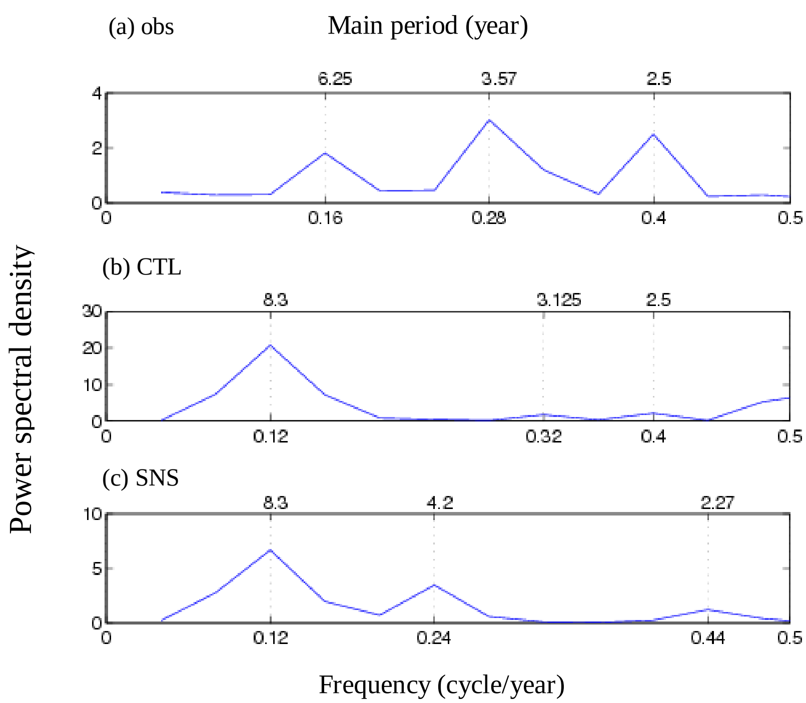

Figure 8. Power spectrum estimate on DJF SLP anomaly over North Atlantic (55°W-0°E, 27.5°N-47.5°N) for observation (a), for CTL (b) and SNS (c) ensembles in the1972-1997 period.

For the CTL ensemble the main periods on shorter time scales (2-4 years) are inconsistent with the simulated NAO index (Fig. 8b and 7b). Moreover, the most evident difference between them is the energy density. The power spectral density for the anomaly of the Azores High is largely increased in CTL at the quasi-decadal period. It is also higher than observations (Fig. 8a and b). This large bias indicates that the variation of the Azores High on longer time scale (8.3 years) is overestimated in AGCM simulations. Moreover, the variation of the Azores High in CTL contributes to the variability of the simulated NAO mainly on this time scale and not at the quasi-biennial periods, which seems inconsistent with reality. Therefore, the inconsistency for the variation of the Azores High on the time scales investigated may eventually lead to an incoherent NAO index in CTL with reality (Fig. 6d).

After realistic snow forcing is included (SNS), it is interesting to find that the main periods of the variation for the Azores High are consistent with most periods of the simulated NAO index (Fig. 8c and 7c). This is very similar to the above result in observations. Energy density at the quasi-decadal period decreases under snow forcing (Fig. 8b and 8c), it becomes nearly as at the same level as the power spectral density for observed both NAO index and variation of Azores High (Fig. 8c, 7c and 8a). Such consistency is closely associated with the improved NAO index in SNS. Therefore, the contribution of realistic snow cover to the frequency modulation of the NAO may be through the (frequency) variation of the Azores High.

As to the variation of the Icelandic Low, it is found that the observed main periods of the SLP anomaly over Iceland are inconsistent with those of the observed NAO index except for the 3.5-year period (Fig. 9a and 7a). That means the periodic change from interannual to quasi-decadal scales for the Icelandic Low is not synchronous with that of the NAO index except for the 3.5-year period. At the same time, it is interesting to see that a similar 3-year oscillation of the Icelandic Low appears in both CTL and SNS. The stable 3-year oscillation may contribute to the NAO variability at the same frequency after snow forcing (Fig. 7c). Therefore, the inconsistent frequency with the corresponding observed and simulated NAO index implies that the periodic variation of the Icelandic Low contributes less to the NAO variability than Azores High. Similar result is obtained when the chosen area over Iceland is different (not shown).

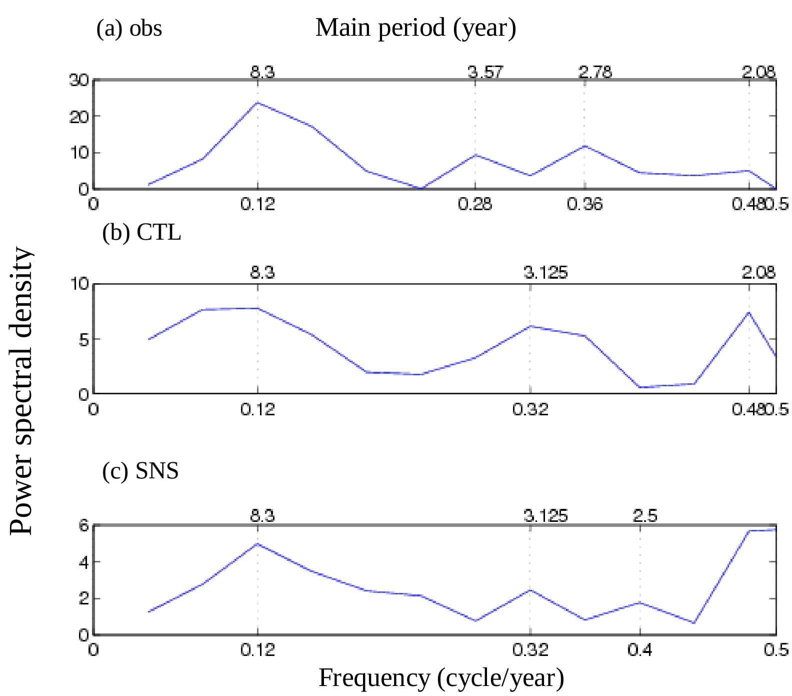

Figure 9. Power spectrum estimate on on DJF SLP anomaly over Iceland (40°W-0°E, 55°N-70°N) for observation (a), for CTL (b) and SNS (c) ensembles in the1972-1997 period.

The atmospheric frequency modulation is a nonlinear process. According to the theory of wave-wave interaction (Ripa, 1981), the modulation effect of snow increases the atmospheric variability when the phases of the oscillations caused by snow and SST are coherent, otherwise the variability in the NAO index is decreased. Studies indicate that the SSTA can modulate the variability of the NAO on longer time scales (Bretherton and Battisti, 2000; Hoerling et al., 2001; Paeth et al., 2003), in this study it is found that the impact of interannual snow variation on the NAO overlaps with the variability cause by SSTA and modulates the variability of the NAO on shorter time scales mainly through the anomaly of the Azores High. Hence, the interannual variability of surface forcing signals in AGCM simulations is very important for the reproduction of the NAO. It should be noted here that many state-of-the-art climate models are to some degree deficient in their ability to simulate the temporal variability of the NAO, especially on shorter (including interannual) time scales (Rodwell et al. 1999; Cassou and Terray 2001; Hoerling et al., 2001; Paeth et al., 2003). This lack of reproduction may partly be caused by the unrealistic interannual variability of surface forcing signals in the models.

5. The effect of interannual snow forcing on NAO on seasonal time scale

The above analysis shows that the interannual snow fluctuations can contribute to modulate the variability of the NAO on shorter time scales (2-4.5 years). Dynamical research on the explanation for actual nonlinear modulation processes at mid- and high latitudes is difficult at present. Due to the less seasonal persistency of snow variation anomalies compared to changes in SSTA, the contribution of snow fluctuations to the variability of the NAO is first of all considered to be important on a seasonal time scale. Statistical studies indicate that autumn Eurasian snow anomalies influence the amplitude of wintertime NAO. In addition, dynamical justification for the influence has also been demonstrated (see Introduction). What corresponding results could be obtained from our simulations on seasonal time scale? Next we will find out how the boundary snow conditions in autumn influence the NAO during subsequent winter.

5.1 Local thermodynamic variations in autumn

The temperature variation at the surface in autumn relates largely with snow through the snow-albedo feedback (see Introduction). It is expected that the observed decreasing trend of Eurasian snow (Fig. 2b) links to seasonal warming at the surface in the same area. The Eurasian warming trend in autumn is well captured in our simulations (Fig. 10a). Especially in the SNS ensemble the increased trend is close to reality. Moreover, the correlation coefficient for the Eurasian surface temperature anomaly between SNS and observations is 0.584, significant at a 99% confidence level (Table 2). The corresponding relationship between CTL and reality is weaker and less significant. The latter was similar to a simulated case study by Watanabe and Nitta (1998) who pointed out that the surface temperature anomaly in eastern Eurasia was moderate and less significant correlated with reality under SST forcing. This means that the local surface temperature variation is largely influenced by change and variability in the local snow cover.

The snow variations lead directly to changes in surface albedo according to the snow parameterization scheme, and thereby influence the local energy budget through radiation processes. The increasing trend appears for the outgoing longwave (LW) radiation flux over Eurasia in both observations and simulations (Fig. 10b), while the decreasing trend takes place for the upward or reflected shortwave (SW) radiation (Fig. 10c). This is reasonable since less snow results in less reflected SW radiation. The remaining absorption of solar radiation in turn warms the surface temperature which leads to increased outgoing LW radiation. This is in agreement with the results of sensitivity experiments with Eurasian snow forcing by Gong et al. (2004).

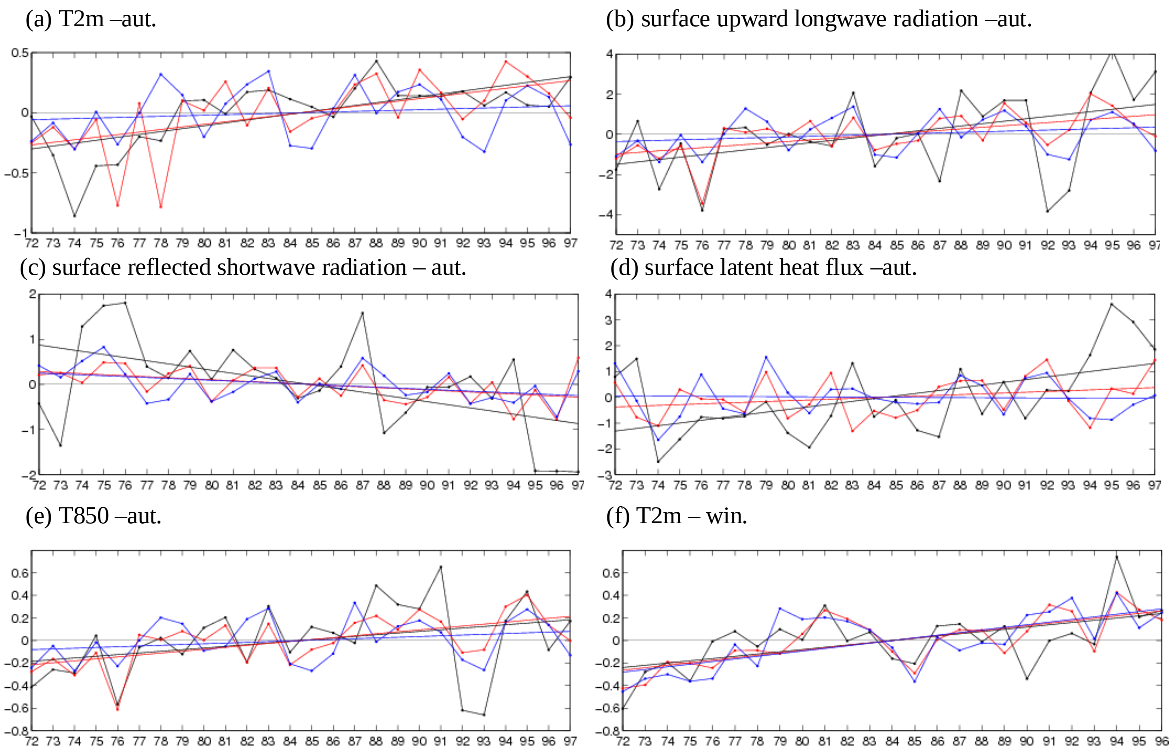

Figure 10: interannual anomalies of area-averaged 2m temperature (a), surface upward longwave radiation (SULWR) (b), surface reflected shortwave radiation (SRSWR) (c), surface latent heat flux (SLHF) (d) and 850hPa temperature over Eurasia (20°N-75 °N, 0-177.5 °E) in autumn and interannual anomalies of area-averaged 2m temperature over Azores High (55°W-0°E, 27.5°N-47.5°N) in winter (f). The temporal mean covers 1972-1997, black line with marks represents the observed variation, blue and red lines are CTL and SNS simulations respectively, the lines without marks show the anomalous trend, the light black line is zero line. Unit: °C and W/m^2.

At the same time the surface latent heat (LH) flux in SNS also increases in time, while no such trend is found in CTL (Fig. 10d). Moreover, in the lower troposphere both the anomaly and the trend in temperature show good consistency with the observations (Fig. 10e), especially for the SNS ensemble. However, the influence is small for temperature in the higher troposphere (not shown). Therefore, increased correlations between energy variables in SNS and observations in Table 2 compared to CTL and observations indicate that the simulated interannual variability of these variables is well improved with realistic snow forcing, except variation of reflected SW radiation at the surface.

	T2m	SULWR	SRSWR	SLHF	T850
CTL	0.2798	**0.556**	**0.368**	0.0931	**0.523**
SNS	**0.5838**	**0.6646**	0.312	0.1256	**0.7385**

Table 2: Correlation coefficients of anomalies for T2m (temperature at 2 meter), SULWR (surface outgoing longwave radiation flux), SRSWR (surface reflected shortwave radiation flux), SLHF (surface latent heat flux) and T850 (temperature at 850hPa) over Eurasia (20°N-75 °N, 0-177.5 °E) in autumn between simulations and observations, the black fonts mean that the correlation is statistically significant at a confidence level higher than 90%.

Because downward shortwave radiation at the surface is not a parameter in the NCEP/NOAA dataset, the clear sky surface downward shortwave radiation has to be used instead. There should be some difference between reflected SW radiation with clouds and clear sky. The simulated variation of this variable with respect to clouds is closer correlated to the variation of observed clear sky reflected SW radiation in CTL than in SNS (Table 2). This maybe implies that the variation of surface reflected SW radiation is not properly represented in CTL. It is just less correlation in SNS than in CTL that shows the difference between these two fluxes. Considering the inconsistency of this flux, the decreased correlation seems reasonable and meaningful.

The thermal influence of snow forcing on the atmosphere is focused to surface and lower troposphere, which is very much in agreement with the results of Kumar and Yang (2003). The above analysis implies that realistic variability of Eurasian snow cover can contribute to simultaneous changes of atmospheric thermal states in the same area at the surface and in lower troposphere on seasonal time scale. The consequence of altered net radiation at the surface could be looked upon as a diabatic heating source or sink in the lower atmosphere.

5.2 Atmospheric variations in winter

The variations in the strength of the Icelandic Low and the Azores High primarily associate with the changes in the surface thermal states in the same areas. For example, the subtropical highs over oceans are deep, warm and stable anticyclonic systems. If the surface temperature anomaly over subtropical North Atlantic is positive, the strength of the Azores High becomes higher than average, and vice versa. Thus, in order to explore seasonal lagged connection between autumn Eurasian snow cover and wintertime NAO, the near surface area-averaged temperature anomalies around the regions of the Icelandic Low (40°W-0°E, 50°N-66°N) and the Azores High (same as above) and their relationship with fluctuations in October Eurasian snow cover are investigated respectively.

Fig. 10f shows wintertime interannual anomalies and trend of the area-averaged 2m temperature over the Azores High. The anomalies in the ensembles show significant correlations to observations: 0.7378 and 0.6808 for SNS and CTL respectively, significant at a 99% confidence level. This implies that the strength of anomalies in the Azores High is largely determined by the persistent SSTA.

It is also found that the thermal variation over the Azores High has a significant negative linear relationship with fluctuations in preceding Eurasian snow cover both for observations and for SNS, but not for CTL (Table 3). On the other hand, no significant connection is found between Eurasian snow cover and subsequent temperature variation over the Icelandic Low. This means that the significant correlation related to the Azores High in SNS is associated with realistic snow fluctuations used in this ensemble. The Eurasian thermal variations in lower troposphere in autumn caused by snow fluctuations will induce subsequent atmospheric response at a larger spatial scale. In this way the interannual snow variation can influence the strength of the Azores High. The negative correlation implies that the more Eurasian snow in October, the colder the wintertime surface air temperature over subtropical North Atlantic and the weaker the Azores High. This contributes to a negative NAO index, and vice versa (see Introduction).

DJF ΔT_{2m} over Azores High	Area-averaged snow cover over Eurasian in Oct.	Correlation coefficient
ERA	ERA	**-0.3783**

SNS CTL	SNS CTL	**-0.5556** 0.0496
DJF ΔT_{2m} over Icelandic Low	Same as above	Same as above
ERA SNS CTL	ERA SNS CTL	-0.2906 -0.0058 0.0089

Table 3: Correlation coefficients between variations of DJF 2m temperature over regions associated with the NAO and autumn Eurasian snow cover in simulations and observation respectively. Black fonts mean that the correlation is statistically significant at a confidence level higher than 90%.

It is noted that the anomalous center of the Azores High in CTL, in the spatial pattern of the NAO (Fig. 6b), shifts toward north and east in SNS (Fig. 6c). Thus, the location and extent of the positive anomalous center in SNS are similar to reality (Fig. 6a). Such eastward shifting for the anomaly of the Azores High also takes place in a case study of 1988/1989 wintertime NAO with deficient autumn Eurasian snow forcing (Watanabe and Nitta, 1998). Furthermore, because of the frequency modulation of snow on the NAO, the consistent phase of the NAO index in SNS with observations after 1982 (Fig. 6d) is another indirect evidence illustrating the role of the Azores High in the variability of the NAO. The reason might be the close relationship between subtropical highs and SSTA, where the boundary forcing expressed by the SST dataset is optimum interpolated after 1982. It is clearly seen that the snow variations contribute to the variability of the NAO on a seasonal time scale through the strength and location anomalies of the Azores High.

6. Summary and discussion

In this paper the influence of the Eurasian snow conditions on the NAO has been investigated in two ensembles AGCM simulations both forced with observed SST. In the control ensemble the snow conditions have been simulated with the model, in the other realistic snow cover variability has been used. The results of the simulations are compared with analyses based on observations.

Both observed and simulated results indicate that the extensive Eurasian snow cover in October heralds the negative wintertime NAO, and vice versa. The negative relationship between autumn Eurasian snow and the NAO index proves the value of seasonal NAO prediction on the basis of snow variation. This influence is not purely a thermal effect of the snow-albedo feedback, dynamic interaction in the atmosphere is important as well (see Introduction).

As we know, the ongoing global warming trend, intensified since around 1980, coincided with a positive tendency in the NAO index (Hoerling et al. 2001; Räisänen 2001). At the same time Eurasian snow anomalies were persistently less than average (Robinson 1996). Many studies have indicated that the autumn snow cover and snow depth anomalies, especially over the Eurasian continent, play an important role in modulating and/or amplifying the variations of NAO through dynamical effects in the atmosphere arising from the snow anomalies (see Introduction). However, snow cover variability represented in AGCMs is far away from the reality. This fact limits evaluation of the contribution of snow variation to extratropical wintertime atmospheric variability simulated by climate models.

By power spectral analysis on the observed and simulated NAO index from 1972 to 1997 in our simulations, it is found that the variability of the NAO on time scales 2-4.5

years is mainly modulated by the realistic interannual snow forcing, while the SSTA seems to dominate the NAO variability on longer time scales. The main periods of the NAO index are correlated with atmospheric anomalies in the Azores High both in observations and in our simulations. On the contrary, there is no evidence for such a consistency for the Icelandic Low. This implies that the variability of the NAO on shorter time scales is modulated by interannual snow forcing mainly through the variation of the Azores High. Further analysis indicates that on seasonal time scale the diabatic heating anomaly caused by Eurasian snow cover in autumn can modulate the strength and location anomalies of the wintertime Azores High by subsequent atmospheric response to the preceding diabatic heating in low troposphere.

However, the mechanism of such a modulation on the NAO caused by snow forcing is still unclear, but must be found by studying atmospheric nonlinear processes, such as wave-wave interaction, wave-mean flow interaction and dispersion/propagation of wave fluxes associated with diabatic heating. The contribution of seasonal snow variations to the NAO through wave-mean flow interaction processes is well studied (Watanabe and Nitta, 1998; Gong et al., 2003). Saito et al. (2001) reported that the strong upward wave flux was spread locally from the troposphere to lower stratosphere in early winter in years with extensive Eurasian snow cover, including the year 1976. In winter of 1976, the case study clearly showed that a zonal wind anomaly in the stratosphere propagated downward to the surface and lead to a wind change southward of 40°N (Black and McDaniel, 2004). Accordingly, it is possible for the strength of the Azores High to vary with the zonal wind anomaly.

On the other hand, Gong et al. (2004) indicated that the produced local temperature response to snow anomalies should be strong enough to modulate the winter AO/NAO. Analysis of our simulations in this paper confirms that the consequence of altered net radiation in autumn after a variation of the interannual snow conditions could be interpreted as a diabatic heating source or sink to the lower atmosphere. The preceding diabatic heating caused by snow variation can modulate the strength and location in anomalies of the wintertime Azores High through atmospheric nonlinear processes.

Randall et al. (1994) reported that the impacts of snow anomalies on the atmosphere largely depend on the AGCMs used. The results in this paper are maybe also tied to the AGCM we have used. Accordingly, our results need to be tested in other AGCMs. At the same time, the dependence of the AGCM implies that the influence of snow conditions on the atmospheric circulation is complex because of the external forcings and the nonlinear dynamics of the atmosphere. In addition, differences between the ensemble with modeled and observed snow cover are partly limited by the internal snow parameterization. A lot of sensitive experiments have been carried out to study the influence of snow on atmospheric thermal states, summer monsoon and wintertime extratropical circulation variability (see Introduction). This study tells us that realistic snow variability is very important for the changes of all the above phenomena. Accordingly, the snow parameterization in AGCMs needs to be improved further.

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