LONG-TERM CHANGES IN COVARIABILITY BETWEEN CONTINENTAL-SCALE SNOW COVER AND ATMOSPHERIC CIRCULATION IN THE NORTHERN HEMISPHERE

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ABSTRACT

Large-scale snow cover extent and the atmospheric circulation have recently been recognized to interact mutually. Understanding the snow-atmosphere interaction is important for a better predictability of the cryospheric climate variability under the future climate change.

We investigated interactions between continental snow cover extent and the large-scale atmospheric circulation variability in the Northern Hemisphere, on a climatic scale (from seasonal to sub-decadal) for the recent three decades. Both observations and numerical simulations were used. Utilized snow cover data were derived from the visible sensor (AVHRR). The indices to present the atmospheric circulation variability, such as the Arctic Oscillation (AO), were computed from the re-analysis data. Statistical analyses showed that the Eurasian snow cover extent (SCE_{EUR}) and the AO have a significant coherency in sub-decadal periods, with the former leading the latter by several months. A more detailed inspection revealed a climatological change in the sub-decadal covariability between snow and the atmosphere in the mid 1980s. The change may be attributed to changes in the seasonal persistency of $SCE_{EUR,}$ from autumn to winter, and in strength of winter-to-spring interaction between SCE_{EUR} and the AO. A series of numerical experiments was performed with an atmospheric general circulation model, to examine the importance of initial (January) circulation anomaly patterns for the change in the winter-to-spring interaction, and reproduced the observed change successfully.

Re-calculation of large-scale snow data using a physical snow model for the pre-satellite era has been planned to reconstruct a dataset of a longer period with consistent quality for an extension of the covariability research. Our preliminary attempt showed, among other atmospheric forcing inputs, the accuracy of the surface air temperatures is most important for a reliable re-construction.

Keywords: Large-scale snow cover, land-atmosphere interactions, climate variability.

INTRODUCTION

Snow cover extent over land (*SCE*), especially on the continental or larger scale, and the atmosphere are interacting mutually. It is widely acknowledged that the atmosphere controls the snow distribution largely through precipitation and temperatures. As opposed to this, it has been recently recognized that snow cover can modify the atmospheric circulation on seasonal to decadal scales, by exerting anomalous thermodynamic forcing to the atmosphere at the lower boundary. The snow-atmosphere interaction is an important piece in understanding the cryospheric climate variability and its potential changes in the future. To improve the predictability of the future change, it is necessary to understand its mechanisms. It will also provide useful information for the planning and management of water resources under the climate change.

Simultaneous and/or lagged seasonal correlations between continental *SCE* and the atmospheric teleconnection patterns were partly recognized by the end of the past century (e.g.1,2). Of the atmospheric teleconnection patterns, the most significant are the Arctic Oscillation (AO) (3) and the North Atlantic Oscillation (NAO) (4). The AO (or NAO) has a structure of a seesaw pattern; one center of action in the high latitude near the polar cap and other centers of action of the opposite polarity along the middle latitude belt. It has been shown that the AO/NAO is the largest source of year-to-year variability of the extratropical boreal winter climate, e.g. surface temperature, storm track, precipitation, and cloud cover (3,4). The *SCE* variability in Eurasia and North America was found to have a statistically significant correlation with the above atmospheric circulation variability, on the climatic scale (from seasonal to sub-decadal), for 1971-2002 (5,6), and there has been a change in cross-seasonal relationship between snow and the atmosphere in the mid-1980s (7). In the first subsection of RESULT, we will show the key results of (7) for an extended analysis period. The second subsection in RESULT will show an attempt with an atmospheric general circulation model (A-GCM) to examine the role of the January pressure anomaly pattern in deriving the difference observed in spring before and after the mid 1980s.

The above *climatological* interaction between large-scale snow cover and the atmospheric circulation has only been examined for the period the satellite observation has been available. To ensure the stability (or periodicity) of the phenomena and to understand the relevant background processes, analyses that cover longer time periods are essential. Datasets of snow states (coverage, depth, and/or mass), with coherent quality and of long and large coverage, will therefore be an important asset to our research. Especially those datasets covering the pre-satellite era are required to examine the past climate. There have been several attempts on statistical reconstruction of continental to hemispheric-scale snow depth/coverage datasets extending back to the pre-satellite era. Some of them date back to the beginning of the past century (e.g. 8,9). Re-calculation of snow state data with physical numerical models is another way. Land surface schemes/models may be employed to calculate the snow states on a large scale in both space and time, where observation-based hydro-meteorological data can be used as atmospheric forcings, i.e. inputs to the model (10,11,12). A common tendency of overestimated duration of snow cover was noticed among many of those model outputs. Biases in forcing data, e.g. near-surface air temperature (T_{2m}) and precipitation, have been suspected as a cause of such discrepancy between the observed and calculated values, apart from uncertainties and deficiencies in land surface models/schemes. We conducted a preliminary test for our attempt to re-construct continental-scale snow state data with a physical snow model and available observational (gridded and/or station) data. The used input data and the employed snow model are explained in the next METHODS section, and the results are shown in the third subsection in RESULTS.

List of abbreviations

A-GCM	Atmospheric general circulation model	ERA15	ECMWF 15 Years Re-Analysis	NIES	National Institute of Environmental Studies (Japan)
AO	Arctic Oscillation	FRCGC	Frontier Research Center for Global Change	SCD	Longest duration of snow cover deeper than 3 cm
AVHRR	Advanced Very High Resolution Radiometer	ISLSCP	International Satellite Land Surface Climatology Project	SCE_{EUR}	Eurasian Snow Cover Extent
CCSR	Center for Climate System Research	NAO	North Atlantic Oscillation	T_{2m}	Surface air temperature at 2 m
CRU	Climate Research Unit	NCAR	National Center for Atmospheric Research (USA)	UDEL	University of Delaware
ECMWF	European Centre for Medium-Range Weather Forecasts	NCEP	National Centers for Environmental Prediction (USA)		

METHODS

Satellite-observed snow cover

For the analysis of the satellite-observed period, we used the snow cover data derived from the visible sensor (Advanced Very High Resolution Radiometer; AVHRR) on the NOAA satellites (13). As the atmospheric variability indices, we use the Arctic Oscillation (AO) and the North Atlantic Oscillation (NAO), that are downloadable from (14) and (15), respectively. For the numerical experiment, we utilized an Atmospheric General Circulation Model (CCSR/NIES/FRCGC A-GCM), an atmospheric component of the coupled GCM developed jointly by Center for Climate System Research, University of Tokyo; National Institute of Environmental Sciences; and Frontier Research Center for Global Change, Japan Agency for Marine-Earth Science and Technology (16). Experiment design will be explained in the RESULT section, prior to the description of the obtained results.

Re-construction of large-scale snow

For the preliminary test of our snow re-construction, we chose a one-year period from September 1987 to August 1988, because both the satellite and station observations are available for the period, and, hence, can be compared to each other for validation.

The physical snow model we employ is a one-dimensional land surface model adaptive to intensely cold regions (17), which participated in the Snow Models Intercomparison Project (18). The model prognostically calculates the energy and water fluxes between atmosphere, biosphere (although we neglected vegetation in this study for simplicity), land surface and soil layers, as well as snow water equivalent and depth when applicable. Hourly forcing values of six hydro-meteorological variables are required as atmospheric forcings, i.e. surface air temperature at 2 m (T_{2m}), humidity, precipitation, wind, and downward short- and long-wave radiation. Since we expect to use gridded reanalysis data as a basis of the atmospheric forcings in our final product, we used gridded data in this study, too. Data taken from International Satellite Land Surface Climatology Project Initiative I dataset (ISLSCP) (19) was used, instead of common reanalysis data, for this dataset provides all of the six forcing variables with 6-hourly temporal resolution for the period. Surface air temperatures of ISLSCP were derived from the operational analysis by the European Centre for Medium-Range Forecast (ECMWF). When results are compared to the station data (96 stations in the former Soviet Union and 235 stations in the USA were used), the gridded values were interpolated to each station, according to an inverse distance weighing method. The maximum search distance is 200 km for grid size of 1.0 degree or smaller, and 300 km if the grid size is 2.5 degrees or larger. As shown below, however, the ISLSCP dataset is found to have large cold biases, especially in the cold season. Hence, we also examined the sensitivity of calculated snow state to the surface air temperature with and without adjustment to the station-based values.

Station data of T_{2m} and snow depth used for adjustment and/or comparison were taken from the following data sets. '*Daily Temperature and Precipitation Data for 223 former U.S.S.R. Stations*' (20), and '*Historical Soviet Daily Snow Depth version 2.0*' (21) which covers 283 stations in the former Soviet Union. The dataset used for the USA stations was '*United States Historical Climatology Network Daily Temperature, Precipitation, and Snow Data for 1871-1997*' (22) which has 1,062 stations. We chose 96 and 235 stations from the former Soviet Union and the USA, respectively, that contain well-quality-controlled snow depth records. The following gridded climatological datasets were also used for comparison of T_{2m}: temperature datasets prepared by University of Delaware (23) and by Climatic Research Unit at University of East Anglia (CRU05) (24), both of which are more directly related to the station observations than the following reanalysis data as illustrated in the RESULT section. The reanalysis datasets ECMWF/ ERA15 (25) and NCEP/NCAR (26) are more model- dependent.

The initial condition (on September 1, 1987) was set equal at all stations, that is, no snow cover and the constant volumetric moisture of 0.1 in all soil layers. A uniform soil temperature profile was also assumed for all stations, set to the value 10 K above the annual mean T_{2m} of the station, based on a preparatory analysis of the former Soviet Union soil temperature profile (27). Impacts of the initial value of the soil temperature (0 K, 5 K, and 10 K above the annual mean T_{2m}) were indiscernible in this study (not shown).

RESULTS

Covariability between large-scale snow and the atmosphere

Figure 1 shows the result of the analysis on seasonal persistency of Eurasian snow cover (Figure 1a) and strength of the inter-seasonal snow-atmosphere connection (Figure 1b), as extended from Saito, Yasunari and Cohen (7). These inter-seasonal connections between snow and the atmosphere are the basis of the sub-decadal covariability. Both results show a returning tendency in recent years to what was observed before the mid-1980s, hinting at a modulation of the inter-seasonal connection at a decadal frequency rather than at a secular trend.

The decadal changes in seasonal correlations (both in the persistency and in the connection) may be closely related to the change in the October SCE_{EUR} trend (Figure 1c). October has been noticed as the key month for the initiation of the autumn-winter linkage between Eurasian snow and the winter NH weather regime (2,5). The correspondence of decreasing (increasing) trend in October SCE_{EUR} and the persistency change may imply a physical/dynamical link between them, but a detailed analysis is left to future research.

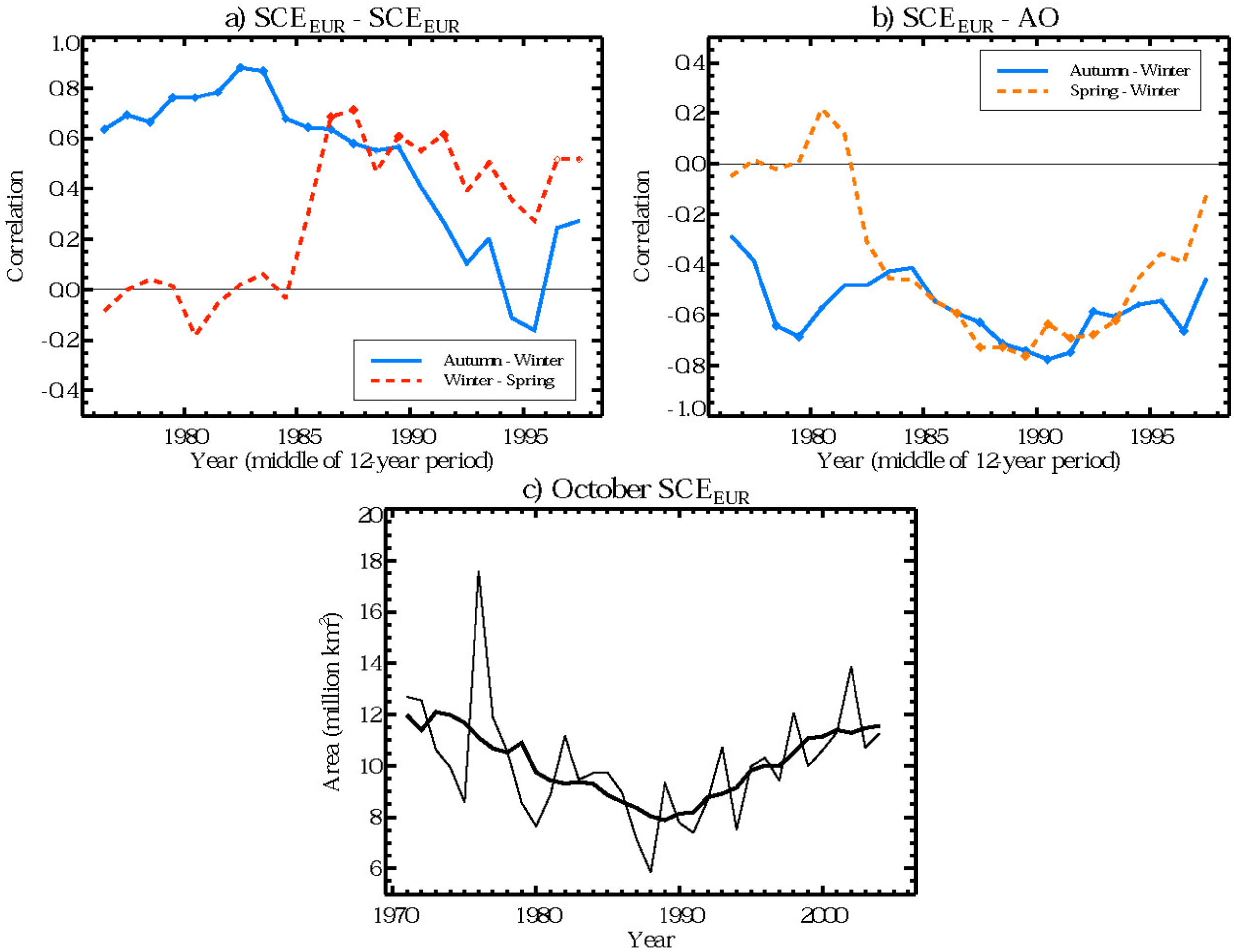

Figure 1: a) Changes in seasonal persistency of Eurasian snow cover (SCE_{EUR}*) between autumn-winter (blue solid), and winter-spring (red dashed) for moving 12-year range from 1972 to 2004. b) As in a) but for strength of inter-seasonal connection between autumn* SCE_{EUR} *and winter AO (blue solid) and between winter AO and spring snow cover (orange dashed). c) Time series of October* SCE_{EUR} *(thin) with its 7-year running mean (thick).*

Examination of changes in winter-to-spring connection by A-GCM

The dynamical mechanism of the autumn-winter connection (Figure 1a) was already examined successfully with the observational evidences, bridging the autumn land surface conditions via the stratosphere polar vortex to the winter tropospheric circulation anomalies (5). As to the causal mechanism of changes in the winter-to-spring connection (Figure 1b), it was only speculated in the previous study that the different anomaly pattern of sea level pressure (SLP) associated with the winter AO/NAO (Figure 2a,b) may result in different thermal advection in spring, and that positive feedback between snow, air temperature, and pressure may maintain the anomaly pattern. Zonal advection was pronounced before the change (Figure 2a), while meridional air from the southern (or northern depending upon the polarity of the AO/NAO) North Atlantic was advected to the continent after the change (Figure 2b). Therefore, in positive AO winters (i.e. lower than average pressure over the Arctic region) comparatively warm air over the Gulf Stream is advected to western Eurasia, which enhances snow melting in the area. Snow-melted land surface is an anomalous heat source to the atmosphere because of lower albedo of the bare (or vegetated) surface, which works to keep the low-pressure anomaly, which in turn brings warm air from the North Atlantic to the continent.

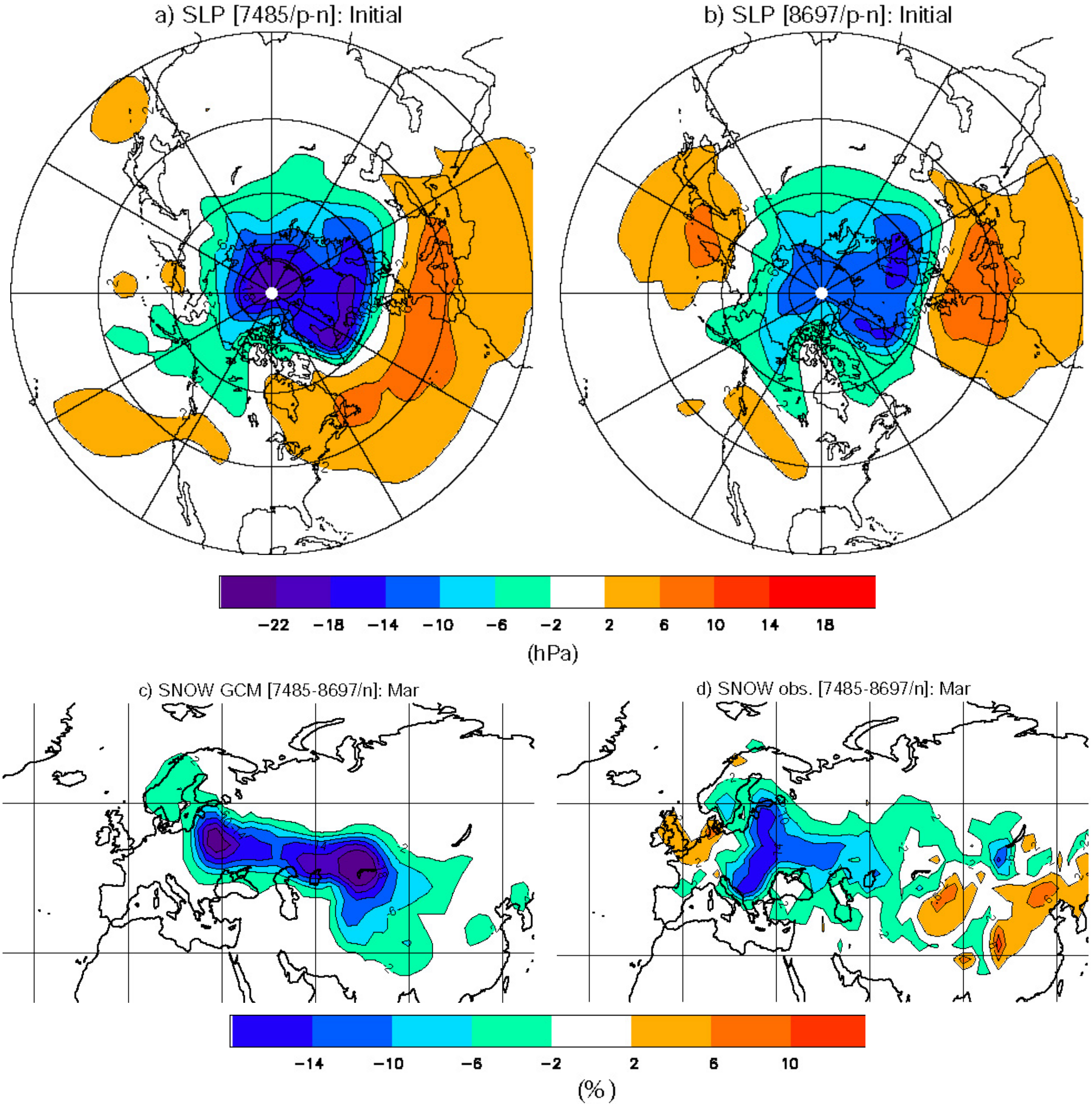

Figure 2: a) Initial anomaly patterns of sea-level pressure (SLP) with respect to the AO-related variations on 1st January for the period from 1974-85. Constructed from the reanalysis data. b) As in a) but for the period from1986-97. c) Difference in March snow cover extent between two periods 1974-85 and 1986-97 following negative-AO winters for model simulation. d) As in c) but for satellite observation.

We performed a numerical experiment using the CCSR/NIES/FRCGC A-GCM to examine the above hypothesis. We conducted four sets of simulations, each starting with four different initial atmospheric conditions on 1st January but other conditions kept equal including initial snow cover extent. The four initial atmospheric conditions are constructed such that they load either *negative* or *positive* anomaly patterns associated with the AO either of *the 1974-85 period* or *the 1986-97 period*. That is, experiments started with

1) 1974-85 *SLP* mean (not shown) plus 1974-85 anomaly (Figure 2a),
2) 1974-85 SLP mean minus 1974-85 anomaly,
3) 1986-97 SLP mean (not shown) plus 1986-97 anomaly (Figure 2b), and
4) 1986-97 SLP mean plus 1986-97 anomaly.

We conducted ten-member ensemble runs for each experiment through the end of May. The result is shown in Figure 2c. It successfully reproduces the anomalously disappearing snow cover in March in western Eurasia following the positive-AO winters for 1986-97, as compared to the observation (Figure 2d).

This experiment, however, only indicates the influence of the given different initial atmospheric conditions upon the evolution of the snow-atmosphere system afterwards. We will need to investigate the cause of the initial pressure anomaly pattern difference before and after the mid-1980s, and also possible influences from prior snow anomaly.

Snow cover reconstruction for the pre-satellite era

As stated in the METHODS section, cold biases were noticed in the ISLSCP and reanalysis data compared to the station-based observations. Figure 3 shows the monthly value of station-wise difference of T_{2m} to illustrate the magnitude and season of the difference among the datasets (abbreviations defined before, i.e. ISLSCP, NCEP/NCAR, ECMWF/ERA15, UDEL and CRU05, are used) averaged over the former Soviet Union and USA regions, respectively.

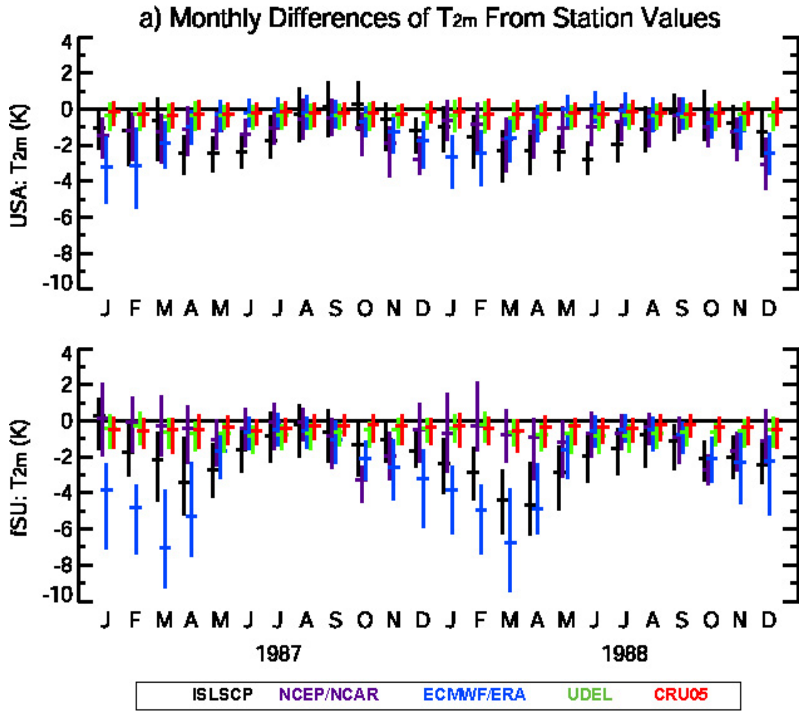

Figure 3: Monthly average of station-wise difference over the USA (upper) and the former Soviet Union (fSU, lower) region for near surface air temperature, T_{2m}. *For abbreviation of the datasets see the text.*

The ISLSCP and the ECMWF reanalysis datasets show a cold bias and large variations between stations in cold seasons in T_{2m}, in both the regions. Examination of possible influences on the snow calculation due to this cold bias motivated us to conduct a sensitivity study (11). Other datasets also show similar cold bias, although much less in magnitude and inter-station variations. Both the UDEL and the CRU05 data are constantly close to the station values, because they are constructed almost directly from station observations.

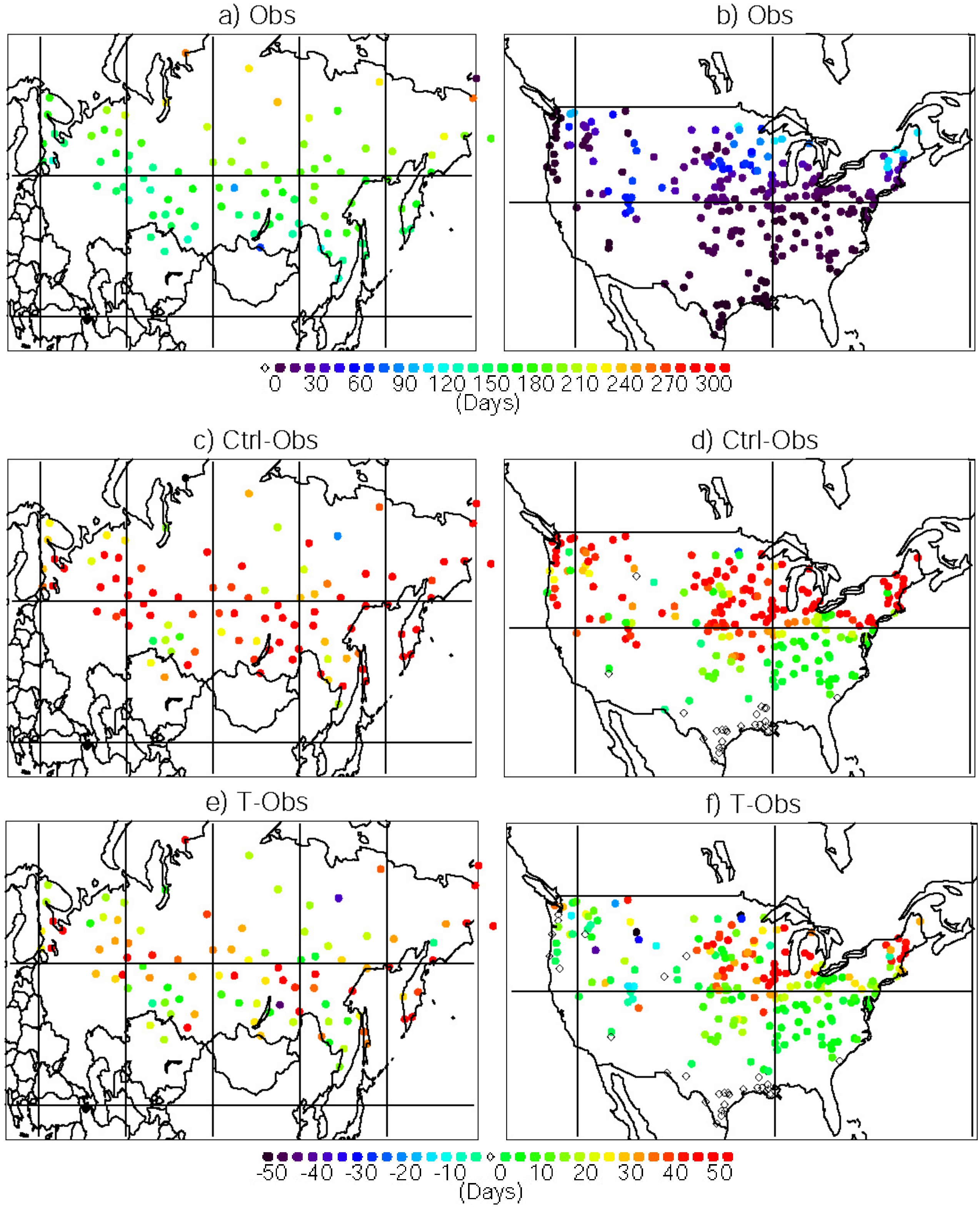

*Figure 4: a) The observed "longest duration of snow cover deeper than 3 cm" (SCD) at stations in the former Soviet Union. c) The difference of the **Ctrl** calculation of SCD from observation. e) As in c) but for **T**. b), d) and f) As in a), c) and e) but for USA.*

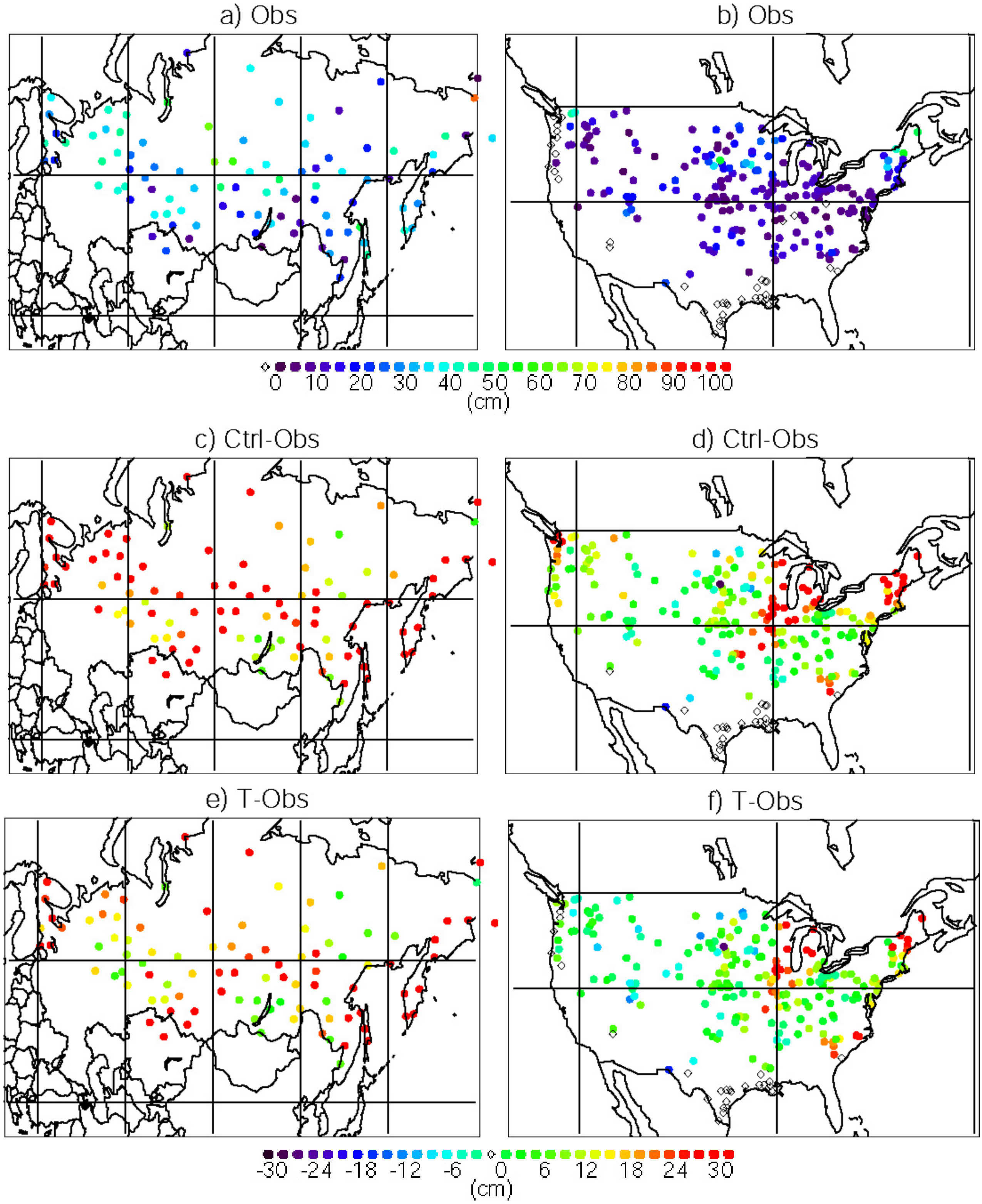

Figure 5: Same as Figure 4 except for mean snow depth during the period with snow cover deeper than 3 cm.

For each of 96 stations in the former Soviet Union and 235 stations in the USA region, we prepared a forcing dataset (set ***Ctrl***) in that the values were interpolated from the ISLSCP for all six variables. Thereby, hourly data were constructed from 6-hourly data in the following way: a) by assumption of a uniform distribution [precipitation], b) by proportional allotment to the theoretically computed values while preserving the total 6-hourly amount [downward solar radiation], and c) by linear interpolation [the other four variables]. Subsequently, we prepared another forcing dataset to examine the sensitivity of the model output with respect to surface air temperature, T_{2m}. The examined set ***T*** was prepared by translating daily T_{2m} to have the same daily mean as the corresponding station value whereby the other five variables were kept unchanged. We then compared the results between ***Ctrl*** and ***T***. Our ‘ground truth’ snow cover data, obtained from two different sources, i.e. station data and interpolated visible satellite data, appeared consistent in both regions (not shown).

Figure 4 shows the result of the T_{2m} sensitivity in terms of the longest duration of snow cover deeper than 3 cm (*SCD*). Substantially similar results are obtained for the distribution of start and end day of *SCD* (not shown), and the mean snow depth during *SCD* (Figure 5).

At most of the stations in the former Soviet Union region, snow cover duration is overestimated by up to 50 days (Figure 4c). Southern stations are relatively closer to the observed value than the northern or inner stations. Adjustment of the input surface air temperatures tends to improve the estimates at stations in tundra region (Figure 4e), although the connection to the vegetation was not clear. Even after the temperature adjustment, there are still stations with overestimations of more than 40 days. At some of those stations the overestimation was due to too early start of snow accumulation, while at other stations the delay of snow melt is responsible due to overestimated accumulation (Figures 5c,d). At some stations, the estimated duration was longer both before and after the observed duration.

In the USA region, estimation of *SCD* was closer to the observed, in comparison to the former Soviet Union region (Figure 4d). Exceptions are stations around the Great Lakes and in the north-eastern and south-eastern areas. Adjustment of the temperature inputs remarkably improved the estimates west of the Rocky regions (Figure 4f). Improvement by the adjustment is more apparent in the mean snow depth field (Figures 5d,f) than in *SCD*.

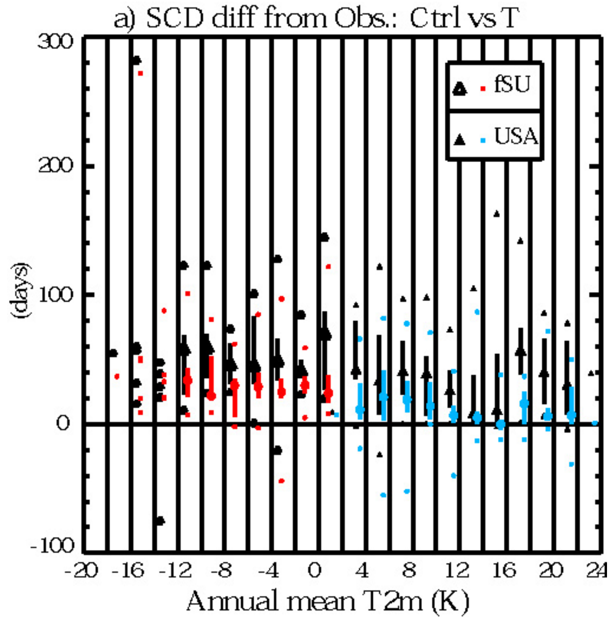

*Figure 6: Difference of snow cover duration between **Ctrl** and **T** averaged over stations stratified by mean annual T_{2m}. Large triangle with black line denotes median and 25%-75% range for fSU stations in **Ctrl**. Large red dot with red line denotes the same but for **T**. Small triangle with black line and blue dot with blue line denote same but for USA stations. Small triangles and dots denote extremes, or individual values if total station number in a lot is smaller than 5.*

It is evident from Figures 4 through 5 that the surface air temperature correction can improve the calculated snow state in both regions, both in mean and cross-station variations. Stratification by mean annual T_{2m} of the stations summarizes the results of the experiment that the estimates are improved in terms of both median and across-station deviations, and that improvement is pronounced in those areas where annual mean T_{2m} is between –12°C and 12°C (Figure 6). It implies that accurate temperature inputs are important for snow calculation, especially when surface temperatures are close to the zero degree, to determine the start of persistent snow cover in the early cold season, or the end of snow melting in the early warm season.

We repeated similar sensitivity experiments with respect to precipitation, but precipitation adjustment showed only a small influence on the snow calculation. This may be explained by the fact that snow cover does not change so much at the beginning of the season, especially in high-latitude areas. The snow model over-estimated the snow depth and snow cover duration even after temperature and precipitation adjustments, although this is what many of the physical snow models experience. This may partly be due to unrealistic initial and vegetation conditions, but improvement of the model physics is also necessary before starting the full re-construction project.

CONCLUSIONS

Extended analysis of the hemispheric-scale snow-atmosphere covariability structure implies a decadal modulation/oscillation rather than a secular change both for autumn-winter and winter-to-spring connections. Our understanding of the dynamical and physical mechanisms needs to be improved. Extending the analysis period and skilful utilisation of a GCM with the fine physical schemes are essential to future research.

The GCM experiment conducted to examine the influence of different initial atmospheric anomaly patterns in winter upon snow cover retreat in the following spring, successfully reproduced the different behaviours before and after the mid-1980s. We plan to further explore the influence of different combinations of initial conditions of the atmosphere and snow cover.

As a part of efforts to extend the temporal coverage of the large-scale, coherent snow cover data, we conducted a preliminary sensitivity test of snow state calculation on input forcing data. The result emphasized the importance of input data quality, especially with respect to surface air temperature; adjustment of T_{2m} improved hindcast skill markedly, especially in those areas where the period of surface air temperature close to zero degree centigrade is relatively long. This implies the necessity and importance of careful *screening* and *pre-processing* of the input atmospheric forcing values prior to the calculation, for which wide collection of quality-assured station-based observational data is inevitable.

The future work on the snow model includes investigations on sensitivity to other forcings (e.g. incoming radiation) as well as inclusion of more elaborate local features such as vegetation and initial soil and snow conditions that will require further observational evidences for validation.

ACKNOWLEDGEMENTS

The authors would like to thank two anonymous reviewers for improving the manuscript.

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