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A longer climate memory carried by soil freeze–thaw processes in Siberia

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Abstract

The climate memory of a land surface generally persists for only a few months, but analysis of surface meteorological data revealed a longer-term climate memory carried by soil freeze–thaw processes in Siberia. Surface temperature variability during the snowmelt season corresponds reasonably well with that in the summer of the following year, when most stations show a secondary autocorrelation peak. The surface temperature memory is thought to be stored as variations in the amount of snowmelt water held in the soil, and through soil freezing, which emerges as latent heat variations in the near-surface atmosphere during soil thawing approximately one year later. The ground conditions are dry in the longer-term climate memory regions, such as eastern Siberia, where less snow cover (higher surface air temperature) in spring results in less snowmelt water or lower soil moisture in the summer. Consequently, through soil freezing, it will require less latent heat to thaw in the summer of the following year, resulting in higher surface air temperature. In addition to soil moisture and snow cover, soil freeze–thaw processes can also act as agents of climate memory in the near-surface atmosphere.

Keywords: climate memory, soil freeze-thaw processes, permafrost

 Online supplementary data available from stacks.iop.org/ERL/7/045402/mmedia

1. Introduction

A vast area of Siberia is covered by taiga forest on continuous permafrost (permafrost regions roughly correspond to the snow-covered areas in figure 1(a)), although in areas with seasonally thawed soil, snowmelt water and summer rainfall can infiltrate the soil following thawing of the frozen soil in the warm season, and this water is stored as ice throughout the cold season. The permafrost regions coincide closely with the distribution of the nearly homogeneous larch-dominated forest. The taiga forest maintains the permafrost by controlling its seasonal thawing, which in turn maintains the taiga forest by providing sufficient water, forming a

taiga–permafrost coupled system (Shur and Jorgenson 2007, Zhang *et al* 2011). In eastern Siberia, the snow cover tends to be relatively thin, and the maximum snow depth is typically about 40 cm in March (figure 1(b)). The insulating effect of snow cover has a more significant effect on ground temperatures, which reach a maximum when snow is at its optimal thickness (about 40 cm) (Zhang 2005). As a result, changes in air temperature over the permafrost surface may be greatly influenced by changes in the seasonal snow cover in this region, because snow depth during the winter varies significantly around the optimal thickness (figure 1(b)).

Snow cover in eastern Siberia plays a critical role in the land–atmosphere climate system through the seasonal climate memory during the snow-free season (Matsumura *et al* 2010, Matsumura and Yamazaki 2012). The snow-hydrological effect (Barnett *et al* 1989, Yasunari *et al* 1991) is prominent for several months during the summer in eastern Siberia;



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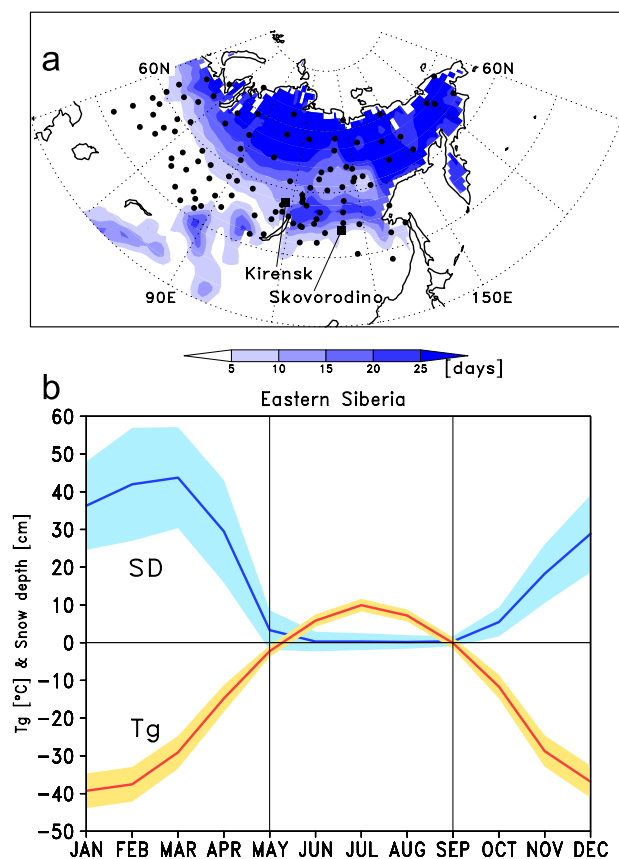


Figure 1. (a) Mean number of days with snow cover in May, as observed by SSM/I for the period 1988–2004 (Japan Meteorological Agency 2005). Black circles and squares denote the locations of observation stations used in this study. Data from Kirensk (57.767°N, 108.067°E) and Skovorodino (54.0°N, 123.967°E) were analyzed in detail. (b) Seasonal changes in mean snow depth (SD: cm) and ground-surface temperature (T_g : °C) with standard deviations (shadings) averaged over the areas 55°–65°N and 100°–130°E for the period 1950–2008.

i.e., enhanced soil moisture (snowmelt water) persists later into the summer, thus contributing to surface cooling, and a coupling between evaporation and precipitation. Less snow cover (higher surface air temperature) in spring results in less snowmelt and correspondingly lower soil moisture in the summer, resulting in turn in reduced cloudiness, evaporation and precipitation (higher surface air temperature). Consequently, the snow-hydrological effect and precipitation recycling maintain the surface heating or cooling during the summer in eastern Siberia, because of the dry ground conditions. Using an atmospheric general circulation model, Delworth and Manabe (1989) assessed the effect of soil moisture on the near-surface atmosphere, and found that soil moisture anomalies persist for only a few months, as measured by autocorrelation. However, in the seasonally frozen subsurface soil of northern Eurasia, soil temperatures show a secondary autocorrelation peak approximately one year later (Schaefer *et al* 2007). The past soil temperature anomalies are stored as variations in the amount of ground ice, and can reemerges at the surface due to thawing of the frozen soil after approximately one year.

We analyze surface meteorological data in an attempt to determine the processes that lead to the development of a longer-term climate memory in Siberia. This longer climate memory does not mean the persistence, but literally the memory. The lack of observational data presents a serious problem when attempting to understand soil moisture in Siberia; consequently, instead of soil moisture, we used ground-surface temperature and surface air temperature as being representative of the land-surface and near-surface atmosphere, respectively.

2. Data and method

We analyzed the Baseline Meteorological Data in Siberia (BMDS) dataset version 5.0, compiled and quality checked by the GAME-Siberia project (Yabuki *et al* 2011). The dataset contains daily mean data for 16 meteorological elements over a 59 yr period (1950–2008). We used the monthly mean of ground surface (soil or snow) temperature (T_g), minimum (T_{min}) and maximum (T_{max}) daily surface air temperature, averaged daily surface air temperature (T_a), and snow depth from 107 stations (figure 1(a)). These data were processed prior to the correlation analysis to remove linear long-term trends, and we used anomaly data by subtracting the average monthly values. For our 59 yr lag data, the time series lost one year, consequently, we subtracted one degree of freedom. We estimated statistical significance using a Student *t*-test with the number of degrees of freedom (56) based on the number of years. To confirm the climate memory, we extended the autocorrelation to northern Eurasia using the National Centers for Environmental Prediction (NCEP) reanalysis (R-2) data (Kanamitsu *et al* 2002), in which the linear trend for the period 1979–2008 was removed.

3. Results and discussion

Before examining the longer-term climate memory, the persistence of the summer soil moisture in eastern Siberia is considered. The persistence associated with soil moisture can be detected more readily in monthly surface temperatures (Walsh *et al* 1985, Delworth and Manabe 1989), and one measure of the temporal variability of monthly surface temperature is the lag-one autocorrelation coefficient, which we computed using data from the months of June, July and August in supplementary figure S1 (available at stacks.iop.org/ERL/7/045402/mmedia). Strong signals of T_g , T_{min} , T_a and T_{max} are seen in eastern Siberia, where the persistence of the summer climate is strong as a result of the seasonal climate memory (Matsumura *et al* 2010, Matsumura and Yamazaki 2012). T_{min} correlations more closely resemble those of T_g than T_a , and are stronger than T_{max} , indicating that the surface climate memory can be detected more easily using T_{min} . Figure 2(a) shows the seasonal changes in correlation between T_g and T_{min} , and between T_g and T_{max} at Kirensk (57.767°N, 108.067°E) and Skovorodino (54.0°N, 123.967°E). The correlation between T_{min} and T_g is high (above 0.8) at both stations and in all seasons. The correlation between T_{max} and T_g is also clearly high during

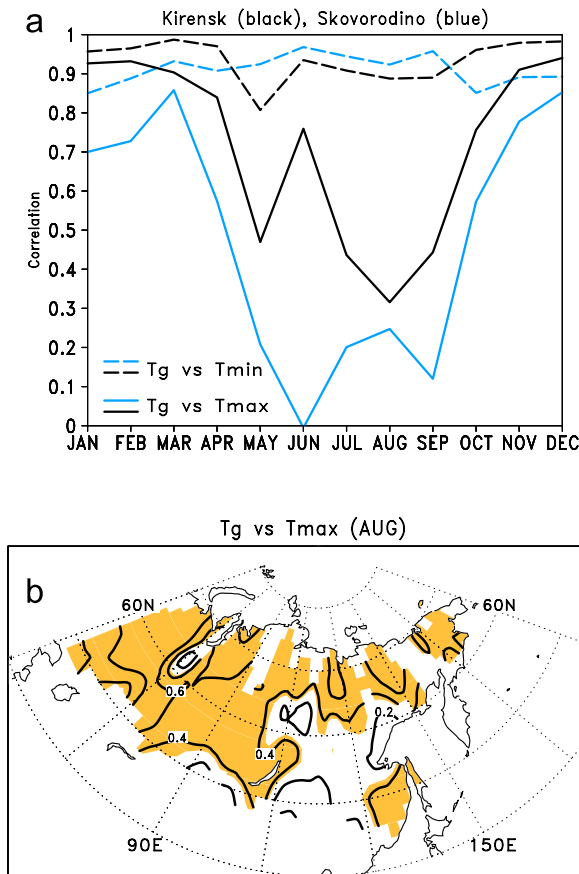


Figure 2. (a) Seasonal changes in correlation between T_g and T_{min} (dashed lines) and between T_g and T_{max} (solid lines) at Kirensk (black) and Skovorodino (blue). (b) Correlation between T_g and T_{max} in August (contour interval is 0.2). Shaded regions are statistically significant at the 1% confidence level.

the snow-covered season, but decreases significantly during the snow-free season. The correlations at Kirensk in August, and at Skovorodino from May through September are not statistically significant at the 1% confidence level. These results suggest that during the snow-free season T_{min} is readily influenced by the ground surface, but T_{max} is not, because daytime cloud cover substantially contributes to T_{max} (Dai *et al* 1999, Tang and Leng 2012). Although the correlation of T_{min} with T_g is high over most of Siberia (not shown), the weaker correlation associated with T_{max} appears over eastern Siberia in August, while over western Siberia T_g is well correlated with T_{max} (figure 2(b)). Thus, surface air temperature in eastern Siberia is more strongly coupled with the land surface than in western Siberia as well, a result consistent with Matsumura *et al* (2010). Accordingly, we discuss climate memory using T_g , T_{min} and T_a in the remainder of this paper.

In eastern Siberia at around 60°N, the snow ablation occurs typically during May, and the soil-thaw season is from May to September (figure 1(b)). Figure 3 shows the time series of lag autocorrelations of T_g and T_{min} upon May at Kirensk and Skovorodino. At Kirensk, the lag autocorrelations between both T_g and T_{min} in May, and those of the following summer are below the significance

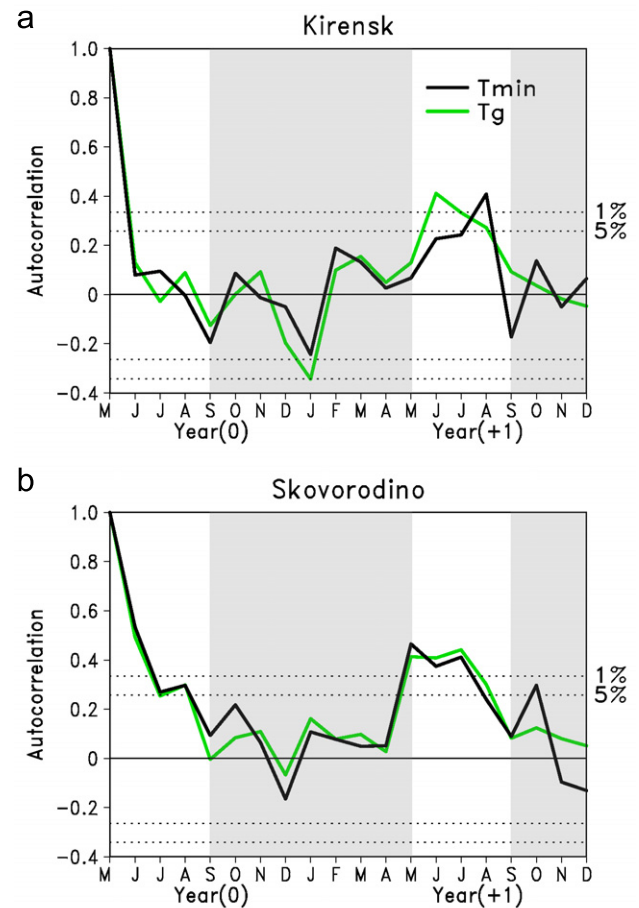


Figure 3. Time series of lag autocorrelations upon May for T_g (green line) and T_{min} (black line) at (a) Kirensk and (b) Skovorodino in the current year (year 0) and following year (year + 1). Dotted lines indicate statistical significance at the 5% and 1% confidence levels, based on the *t*-test. Shadings indicate soil-freeze season in the surface layer.

level, probably due to frequent saturation of the soil layer by snowmelt (Delworth and Manabe 1988). In contrast, at Skovorodino, the correlations of T_g and T_{min} at lags of one (June), two (July) and three (August) months are clearly high, also indicating the persistence of soil moisture through the summer. The difference between the two stations possibly results from the regional differences in the soil thawing and snowmelt seasons, and the resulting meltwater runoff. These correlations decrease in September, when the soil surface begins to freeze, and from then on, during the period when the soil is frozen, they appear to display random variations, although at Kirensk, they are above the significance level in January of the following year. Interestingly, there is a secondary autocorrelation peak in the summer of the following year at both stations. At Kirensk, the correlation of T_g reaches a maximum ($r = 0.41$) in June of the following year, and that of T_{min} reaches a maximum ($r = 0.41$) in August of the following year. At Skovorodino, the correlation of T_g reaches a maximum ($r = 0.44$) in July of the following year, and that of T_{min} reaches a maximum ($r = 0.46$) in May of the following year. Both stations have a secondary autocorrelation peak in the summer of

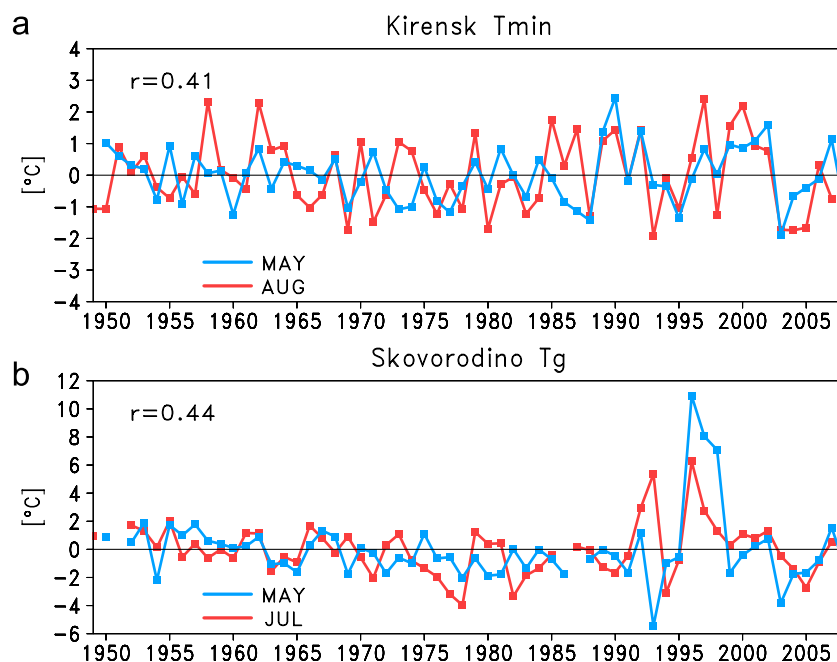


Figure 4. Time series of (a) T_{min} (°C) anomalies at Kirensk in May (blue line) and August (red line; one year shifted for comparison) and (b) T_g (°C) anomalies at Skovorodino in May (blue line) and July (red line; one year shifted for comparison). The linear long-term trends are removed for the comparison of interannual variability.

the following year. These correlations rapidly decrease in September, when the soil surface begins to freeze. Figure 4 shows the time series for T_{min} anomalies at Kirensk in May, and August of the following year (one year shifted in the figure), and T_g anomalies at Skovorodino in May, and July of the following year. The interannual variations of T_g and T_{min} in May correspond reasonably well with those in the summer of the following year at both stations, although at Skovorodino, a strong temperature anomaly in the late 1990s possibly affected the autocorrelation. These variations appear to correspond well before the 1970s, and again from the late 1980s. As most of the temperature data have upward trends, the lag correlations increase if long-term trends are not removed (not shown). There are no lag autocorrelations of T_{max} at either stations (not shown), similar to the results in figure 2.

Other stations in central and eastern Siberia also display a high correlation with the summer of the following year. Figure 5 shows autocorrelations maps of T_g , T_{min} , and T_a between May, and the summer of the following year. Strong signals occur over central Siberia in June of the following year, corresponding to the nearby limit of the snow-covered area in May (figure 1(a)), while in August of the following year they occur mainly over eastern Siberia. In particular, T_{min} correlations occur widely over eastern Siberia. In addition to the snowmelt season, the soil thawing season in central Siberia is earlier than that in eastern Siberia. Thus, strong temperature signals appear to occur over central Siberia in early summer, while over eastern Siberia they occur in late summer. This longer climate memory cannot be explained by the usual surface soil moisture effect (i.e., seasonal climate memory), because it persists for only a few months during the summer. This longer climate memory is probably caused by

the re-emergence of past subsurface soil moisture anomalies stored as variations in the amount of ground ice and latent heat through soil freeze–thaw processes (Schaefer *et al* 2007). Although the presence of seasonal snow cover was not considered by the model of Schaefer *et al* (2007), our results suggest that surface temperature memory is stored as variations in the amount of snowmelt water in the soil, and emerges as latent heat variations in the near-surface atmosphere during soil thawing in the summer of the following year. Higher T_a (less snow cover) in spring therefore results in less snowmelt water and correspondingly lower soil moisture in summer, and through soil freezing, which emerges as less latent heat during soil thawing, and resulting in a higher T_a in the summer of the following year. Schaefer *et al* (2007) also indicated that soil moisture variability in drier soils could produce significant variability in the strength of soil temperature anomaly reemergence. The ground conditions are dry in eastern Siberia, but wet in northwestern Siberia, and ground conditions that are sufficiently dry to enable the storage of snowmelt water within the soil are an important factor in terms of the longer-term climate memory, as well as the seasonal climate memory (Matsumura *et al* 2010). In contrast, in the wet ground conditions in northwestern Siberia, snowmelt water leads to increased runoff; thus, no climate memory is found in northwestern Siberia (see also figure 6).

A secondary autocorrelation peak also appears in southwestern Siberia (figures 5(d)–(f)), where snow ablation occurs mainly in April, although snow remains beyond this time in the high mountainous regions. Because of a lack of data from the regions around 50°N, we confirm the climate memory using NCEP reanalysis data. Figure 6 shows the autocorrelations of T_a from April, May, and June, with August of the following year, from 1979 to 2008. In April, positive

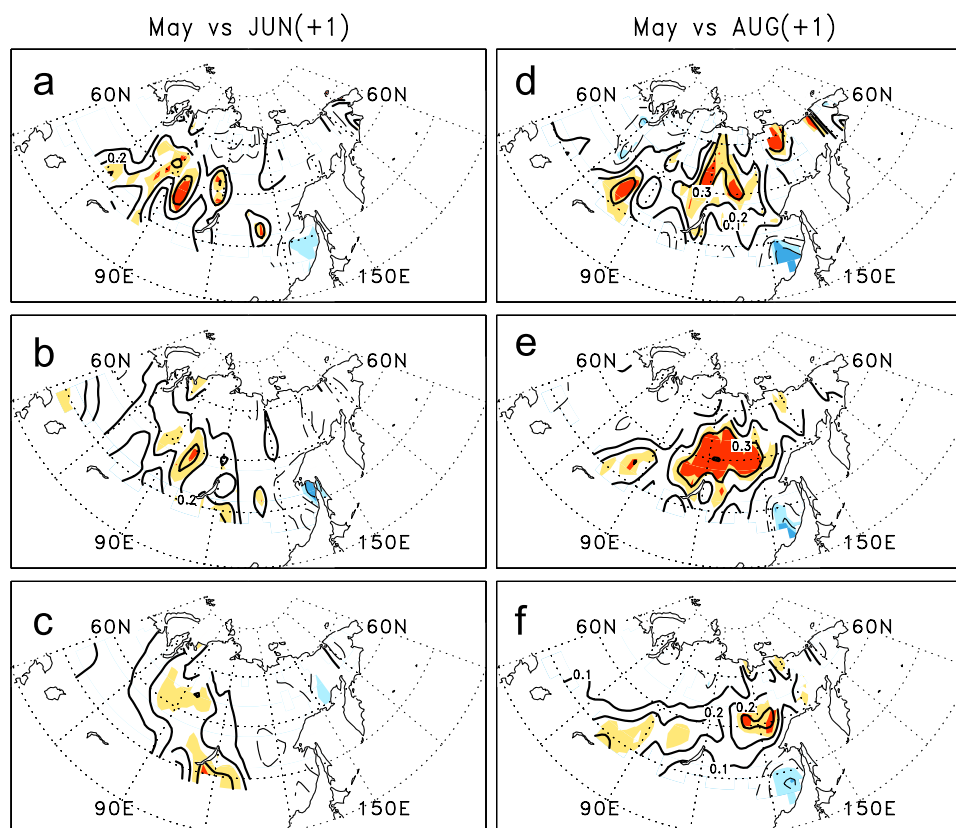


Figure 5. Lag autocorrelations of (a) T_g , (b) T_{min} , and (c) T_a between May and June of the following year. (d)–(f) As for (a)–(c), but between May and August of the following year. Light and heavy shadings represent statistical significance at the 5% and 1% confidence levels, respectively. The contour interval is 0.1, with the zero contour omitted.

T_a correlations are seen in northeastern and southwestern Siberia, where strong T_a signals extend from the Kazakhstan steppe to western Mongolia. As the ground conditions are dry in the Kazakhstan steppe and southwestern Siberia (Vinnikov and Yeserkepova 1991), the longer climate memory is well developed in these regions. On the Kazakhstan steppe, increased snow mass leads to excess soil moisture after snowmelt (Shinoda 2001); thus, seasonal climate memory, which also appears in Mongolia (Yasunari *et al* 1991), may be active in central Eurasia. It is possible that a longer climate memory readily develops in permafrost regions, where meltwater is stored within the active layer. Indeed, southwestern Siberia and western Mongolia contain large areas of permafrost. Strong T_a correlations in May are similar to figure 5(f), and those in June occur over southeastern Siberia, where snow can remain in June because of high mountainous region. The distribution of these longer-term climate memories may reflect the timing of snowmelt and soil thawing seasons, as shown in figure 5.

The seasonal climate memory in eastern Siberia is caused by the soil moisture of the shallow soil layer (Matsumura *et al* 2010). *In situ* observations near Yakutsk in the Lena Basin (Sugimoto *et al* 2003) show that the soil moisture in the shallow soil layer usually decreases during summer after an increase of snowmelt water, while in the deeper soil layer, the soil moisture increases during summer as a result of the infiltration of snowmelt water and melting ice. Sugimoto *et al*

(2003) also indicated that any excess water that accumulates during the period just before the soil freezes is stored as ice through the winter, and water stored as ice in the lower part of the active layer has a residence time of at least 2 yr. It appears that the soil moisture stored as ice in deeper soil layers contributes to the longer climate memory as latent heat variations during soil thawing. The impermeability of frozen soil to snowmelt water leads to increased runoff of meltwater during spring. Moreover, because of the insulating effect of snow cover, it can either enhance or reduce soil freeze–thaw processes, depending on the timing, thickness, and physical and thermal properties of the snow cover (Zhang 2005). Further research is required to better understand the processes that lead to the development of the longer-term climate memory in terms of the interaction between snow cover and frozen soil.

The wintertime North Atlantic Oscillation correlates with vegetation activity in northern Asia (Siberia) during the growing season of 1.5 yr later (Wang and You 2004). This lagged response in vegetation activity, such as taiga forest, is possibly related to the longer climate memory found in this study. Snowmelt water plays an important role in vegetation activity (Grippa *et al* 2005), and permafrost maintains the taiga forest by providing sufficient water, such as snowmelt water and summer rainfall (Zhang *et al* 2011). Consequently, we need to treat the cryospheric climate at the land surface as a snow–permafrost–taiga coupled system. The longer-term

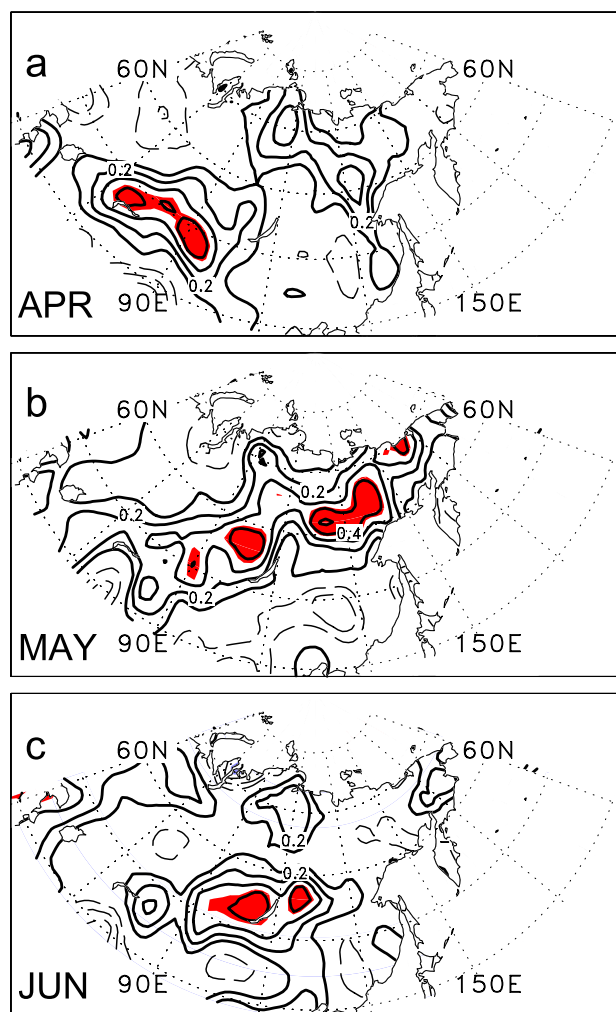


Figure 6. Lag autocorrelations of surface air temperature from NCEP reanalysis data (2 m air temperature) from 1979 to 2008 between (a) April, (b) May, and (c) June and August of the following year. Shaded regions are statistically significant at the 5% confidence level. The contour interval is 0.1, with the zero contour omitted.

climate memory may provide evidence that soil thawing maintains the taiga forest activity in dry ground conditions on permafrost.

4. Conclusions

We analyzed surface meteorological data in an attempt to determine the processes that lead to the development of a longer-term climate memory in Siberia. Surface temperature variability in the snowmelt season corresponds reasonably well with that in the summer of the following year, when most stations show a secondary autocorrelation peak. This longer climate memory cannot be explained by the usual surface soil moisture effect, because it persists for only a few months. The surface temperature memory is probably stored as variations in the amount of snowmelt water held in the soil which, through soil freezing, emerges as latent heat variations in the near-surface atmosphere during soil thawing in the following year. The ground conditions are dry

in the longer-term climate memory regions, such as eastern Siberia, where less snow cover (higher T_a) in spring results in less snowmelt water or lower soil moisture in the summer. Consequently, through soil freezing, it will require less latent heat to thaw in the summer of the following year, resulting in higher T_a . Ground conditions that are sufficiently dry to enable the storage of snowmelt water within the soil are an important factor in the development of a longer climate memory. During soil thawing, liquid water and ice coexist in the soil. Under such conditions, it is difficult to distinguish between snow meltwater and ice meltwater using soil moisture data. The longer-term climate memory appears to emerge as latent heat variations through the phase transitions of water and ice; thus, surface air temperature, especially T_{min} , is a useful indicator of surface climate memory. In addition to soil moisture and snow cover, soil freeze–thaw processes can also act as agents of climate memory in the near-surface atmosphere.

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