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Soil temperature in Canada during the twentieth century: complex responses to atmospheric climate change

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Abstract. Most climate records and climate change scenarios projected by general circulation models are for atmospheric conditions. However, permafrost distribution as well as ecological and biogeochemical processes at high latitudes are mainly controlled by soil thermal conditions, which may be affected by atmospheric climate change. In this paper, the changes in soil temperature during the twentieth century in Canada were simulated at 0.5-degree latitude/longitude spatial resolution using a process-based model. The results show that the mean annual soil temperature differed from the mean annual air temperature by -2 to 7 ˚C, with a national average of 2.5 ˚C. Soil temperature generally responded to the forcing of air temperature, but in complex ways. The changes in annual mean soil temperature during the twentieth century differed from that of air temperature by –3 to 3 ˚C from place to place, and the difference was more significant in winter and spring. On average for the whole of Canada, the annual mean soil temperature at 20 cm depth increased by 0.6 ˚C while the annual mean air temperature increased by 1.0 ˚C. Three mechanisms were investigated to explain this differentiation: air temperature change altered the thickness and duration of snow cover, thereby altering the response of soil temperature; seasonal differences in changes of air temperature; and changes in precipitation. The first two mechanisms generally buffer the response of soil temperature to changes in air temperature, while the effect of precipitation is significant and varies with time and space. This complex response of soil temperature to changes in air temperature and precipitation would have significant implications for the impacts of climate change.
KEY WORDS: soil temperature, air temperature, climate change, high latitudes.

1. Introduction

Air temperature at high latitudes has increased at a higher rate than the global mean during the twentieth century [*Serreze et al.,* 2000; *Zhang X. et al.,* 2000], and most general circulation models (GCMs) project that this pattern will continue [*McCarthy et al.,* 2001]. High latitude (boreal and tundra) ecosystems occupy about 22% of the global land surface and contain about 40% of the world's soil carbon [*Post et al.,* 1982; *McGuire et al.,* 1995]. About 24% of the exposed land in the Northern Hemisphere is classified as permafrost regions [*Zhang T. et al.,* 1999]. Changes in soil temperature associated with climate warming may result in thawing of permafrost, changes in terrain and hydrologic conditions, alteration of the distribution and growth rate of vegetation, enhancement of soil organic carbon (SOC) decomposition, and increased CO_2 emissions from the soil to the atmosphere [*Oechel et al.,* 1993; *Trumbore et al.,* 1996; *Goulden et al.,* 1998; *Nelson et al.,* 2001; *Nelson,* 2003]. These effects could have significant consequences locally and globally.

Most climate records and climate change scenarios projected by GCMs are for atmospheric conditions. Soil thermal and hydrological conditions, however, are the major factors determining the distribution of permafrost, and constraining ecological and biogeochemical processes at high latitudes, such as the onset of growing seasons, root activities, SOC decomposition, and nitrogen availability for plants, etc. [*Moore,* 1981; *Long and Woodward,* 1988; *Oechel et al.,* 1993; *Trumbore et al.,* 1996; *Goulden et al.,* 1998]. Soil moisture at high latitudes is also closely related to soil thermal conditions through snow dynamics, seasonal thawing and freezing, and active-layer thickness in permafrost regions. Many features of high latitudes can modify the dynamics of soil temperature: the prolonged and often thick snow cover; widespread organic materials in the soil (especially in peatlands); annual thawing and freezing of the upper soil layers; permafrost and ground ice [*Smith,* 1975; *Goodrich,* 1982; *Smith and Riseborough,* 2002]. For example, soil temperature in winter can be up to 20 °C higher than the air temperature, and the annual mean soil temperature can be 1 - 11 °C higher than the annual mean air temperature due to the insulating effect of snow [*Smith,* 1975; *Smith and Riseborough,* 2002]. A surface organic layer (forest floor or peat) can insulate the soil from warming in summer and generally results in a lower soil temperature [*Smith,* 1975]. Therefore, the annual means and seasonal variations of soil temperature at high latitudes can be significantly different from that of air temperature.

More importantly, the long-term changes in soil temperature may differ from that of air temperature, because changes in vegetation, snow, soil moisture, and other climate variables (i.e., precipitation, solar radiation, and humidity) all can influence water and energy fluxes on the surface and in the soil, and therefore modulate the relationship between soil and air temperatures. Recent studies show that changes in snow depth and precipitation are as important as changes in air temperature to the measured changes in soil temperature [*Pavlov,* 1994; *Schmidt et al.,* 2001; *Zhang T. et al.,* 2001; *Harris et al.,* 2003; *Stieglitz et al.,* 2003]. Model sensitivity analysis shows that changes in site-conditions could induce to large changes in soil temperature [*Smith and Riseborough,* 1983], and surface organic layers could shield the ground thermal regime from the effects of atmospheric climate change [*Riseborough,* 1985]. Therefore we cannot simply apply the projected or measured warming trend in the atmosphere to the ground [*Williams and Smith,* 1989]. Because of the importance of soil temperature to permafrost distribution and ecological and biogeochemical processes at high latitudes, understanding the responses of soil temperature to the changes in atmospheric climate would be the first step to assess the impacts and feedbacks of climate change.

Canada, the major high-latitude region of North America, includes permafrost and non-permafrost areas, and a range of climate, vegetation and soil conditions. Increasingly detailed spatial datasets have been developed through utilization of remote sensing and geographical information systems (GIS) techniques. In this study, the dynamics of the soil thermal regime in Canada during the twentieth century (from 1901 to 1995) was simulated at a spatial resolution of 0.5-degree latitude/longitude. The tempo-spatial patterns of near-surface soil temperature were analyzed in relation to that of air temperature. Possible mechanisms were investigated to explain the difference between the changes in soil and air temperatures.

2. The Model and Input Data

We developed a process-based model of Northern Ecosystem Soil Temperature (NEST) to simulate the transient response of soil thermal regime to climate change [*Zhang Y. et al.,* 2003]. The model explicitly considered the effects of different ground conditions, including vegetation, snow cover, forest floor, peat layers, mineral soils and bedrock (Figure 1). The dynamics of soil temperature were simulated by solving the one-dimensional heat conduction equation, with the upper boundary conditions (the ground surface or snow surface when snow is present) determined based on the energy balance and the lower boundary conditions (at 35 m depth) defined as the geothermal flux. The ground profile was divided into 40 layers and the thickness of the layers gradually increased from 0.1 m for top layers to 2 m close to the lower boundary. The snowpack was also divided into about 0.1 cm layers, and the number of snow layers and the thickness of the snowpack were updated every day based on snow dynamics. Heat capacities of ground and snow layers were calculated from the specific heat capacity of liquid water, ice, organic materials, minerals and air, weighted according to their respective volumetric fractions. The thermal conductivity of each ground layer was calculated as the geometric mean of the thermal conductivities of the constituents. The thermal conductivity of each snow layer was estimated based on its density. The profile of snow density was simulated considering compaction and destructive metamorphism for each layer. The thickness of the snowpack was determined based on snow density and the amount of snow on the ground (water equivalent), which was the accumulative difference between snowfall and snowmelt. Snowmelt, sublimation, and

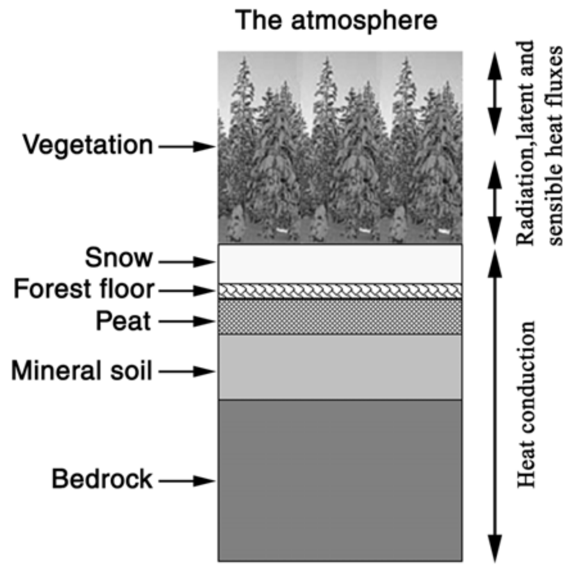

Figure 1. The components of the ground strata and energy fluxes considered in the NEST model.

evapotranspiration were determined based on surface energy balance. Soil water dynamics were simulated considering water input (rainfall and snowmelt), output (evaporation and transpiration) and distribution among soil layers. The effects of thawing or freezing on soil temperature as well as the fractions of ice and liquid water in a soil layer were determined based on energy conservation: latent heat released or absorbed during freezing or thawing equals the amount of heat required or released for the apparent temperature (soil temperature determined by heat conduction equation without considering the thawing/freezing effects) change of the layer. The depth of thawing or freezing front was determined based on the fractions of liquid water and ice in soil layers. Thus, the model integrated the effects of atmospheric climate, vegetation, and ground strata (snow, forest floor, peat layers, mineral soils, and bedrock) on soil thermal dynamics based on energy and water transfer in soil-vegetation-atmosphere systems. The model was validated against measurements of energy fluxes, snow depth, soil temperature and thaw depth. Detailed description and validation of the model can be found in *Zhang Y. et al.* [2003].

Inputs to the model include information about vegetation (land cover types, leaf area index), ground conditions (thickness of organic layers, texture of the mineral soils, SOC content in mineral soils, ground ice content, and the geothermal flux) and atmospheric climate (air temperature, precipitation, solar radiation, vapor pressure, and wind speed). Vegetation types were determined based on the land cover map of Canada derived from the images of the Advanced Very High Resolution Radiometer (AVHRR) [*Cihlar et al.,* 1999]. The original 29 land cover types (excluding water bodies, and ice/snow) were simplified into 5 types: coniferous forest, deciduous forest, mixed forest, crop/grass land, and shrub/tundra. Leaf area index (LAI) and its seasonal variation were derived from AVHRR 10-day composition images [*Chen J.M. et al.,* 2002]. To match the spatial resolution of climate data, the 1 km resolution land cover map and LAI images were aggregated to 0.5-degree latitude/longitude based on the dominant vegetation type in a grid cell for land cover, and based on the average for LAI. Water bodies were excluded in the calculation. Changes in vegetation types and LAI from year to year were not considered due to lack of long-term data. Vegetation related parameters (i.e., height and wood biomass) were estimated based on LAI and vegetation type (Table 1).

Table 1. Model parameters for different vegetation types and conditions.

	Coniferous forest	Deciduous forest	Mixed forest	Crop, grass	Shrub, tundra	Data source*
Height, m	15	15	15	min(0.5LAI,H_{max})		1
Wood biomass, kg m^{-2}		$5.0LAI_{max}+1.7$		0	0	2
Specific leaf weight, kg m^{-2}	0.28	0.10	0.19	0.10	0.10	3
Light extinction coefficient	0.5	0.58	0.54	0.58	0.58	3
Specific water content, kg kg^{-1}	1.0	1.0	1.0	1.0	1.0	4
Canopy emissivity	0.98	0.98	0.98	0.98	0.98	5
Albedo (no snow)	0.10	0.14	0.12	0.20	0.20	6
Albedo (with Snow)	0.10	0.21	0.15	0.50	0.50	6

* Note and references: 1, the heights of the forests were assumed based on the measurements by *Gower et al.* [1997]; H_{max} is the maximum heights; We used 1.0 m for crops and grass, and 1.5 m for shrub and Tundra; 2, LAI_{max} is the maximum LAI in a year (summertime). The relationship for forests was derived from Canada's inventory data and LAI from AVHRR at the end of July; 3, the values for coniferous and deciduous forests were from *Aber et al.* [1996]; An average value was used for mixed forest, and the values of deciduous forest were used for non-forest vegetation types; 4, *Wenger* [1984]; 5, *Oke* [1978]; 6, *Betts and Ball* [1997].

Soil texture, bulk density and organic content were extracted from the soil landscape database of Canada [*Shields et al.,* 1991*; Tarnocai and Lacelle,* 1996]. There are about 15000 polygons in the database covering the whole Canada landmass. Each polygon contains up to 10 components (with area percentage of composition), and each component includes 3 soil layers. We first interpolated these polygon data to 1 km spatial resolution, and then aggregated them to 0.5-degree latitude/longitude grid cells based on the dominant type in a grid cell for soil texture, and the averages for forest floor thickness, bulk density and soil organic carbon content. Excess ground ice often exists in regions with permafrost and was considered during model initialization utilizing the ground ice content presented on the permafrost map of Canada [*Heginbottom et al.,* 1995]. *Pollack et al.* [1993] calculated the geothermal flux for about 250 sites in Canada. These measurements were interpolated to 0.5-degree latitude/longitude resolution for the whole of Canada.

The gridded climate dataset used in this study had a 0.5-degree latitude/longitude spatial resolution globally and a monthly temporal resolution from 1901 to 1995 [*New et al.,* 2000]. It includes monthly means of air temperature, the diurnal range of air temperature, water vapor pressure, cloudiness, monthly total precipitation, and monthly total wet-days. This dataset was interpolated from station measurements considering topographic corrections. We down-scaled this monthly climate dataset to half hourly to accommodate the short time-step (e.g., 15-30 minutes) required by the NEST model. First, daily values were estimated based on monthly data, and then, the diurnal changes were derived from these daily values. Daily means and diurnal ranges of air temperature, vapor pressure, and cloudiness were linearly interpolated from their monthly means, and then modified to consider if the day was a wet-day (precipitation > 0.1 mm) or a dry-day. Diurnal variation of air temperature was estimated using a sine function for daytime and an exponential function for nighttime [*William and Logan,* 1981]. Daily total solar radiation was estimated on the basis of cloudiness [*Penman,* 1948], and then distributed diurnally using a sine function [*Chung and Horton,* 1987]. The diurnal change in vapor pressure was not considered, but vapor pressure deficit varied diurnally because of the changes in air temperature. There is no wind speed in the dataset; a wind speed of 3.0 m s^{-1} was assumed for the simulation based on climate station measurements. The distribution of wet-days within a month was determined as a random distribution, and the amount of precipitation on a wet-day was determined based on an exponential distribution with less frequency for heavier precipitation events [*Richardson,* 1981; *Hann,* 1977]. A detailed description of the down-scaling and its validity for simulating soil temperature were presented by *Chen W. et al.* [2003].

To simulate changes in soil temperature during the twentieth century, the ground thermal and hydrological regimes for each grid cell must be initialized. Since there are no detailed measurements for model initialization, the initial conditions were determined by iteratively running the model for 50-100 years to an equilibrium (changes in annual mean ground temperature was less than 0.001 ^{o}C) corresponding to an initial climate condition. Using this approach, the simulated ground thermal regimes (especially at greater depth) in the early years may not be comparable with measurements because of the equilibrium assumption. However, this approach gives the pure climate change effects for the simulation period, and the results become increasingly realistic with time as the effects of the assumed initial equilibrium condition have less influence. The initial climate was composed of a whole year so that snow dynamics and their effects on soil temperature were included. The initial climate was derived by linearly extrapolating each month's climate variables during 1901-1995 to 1900. Thus, this initial climate was not an estimate for year 1900, but an estimated average climate for the later nineteenth century.

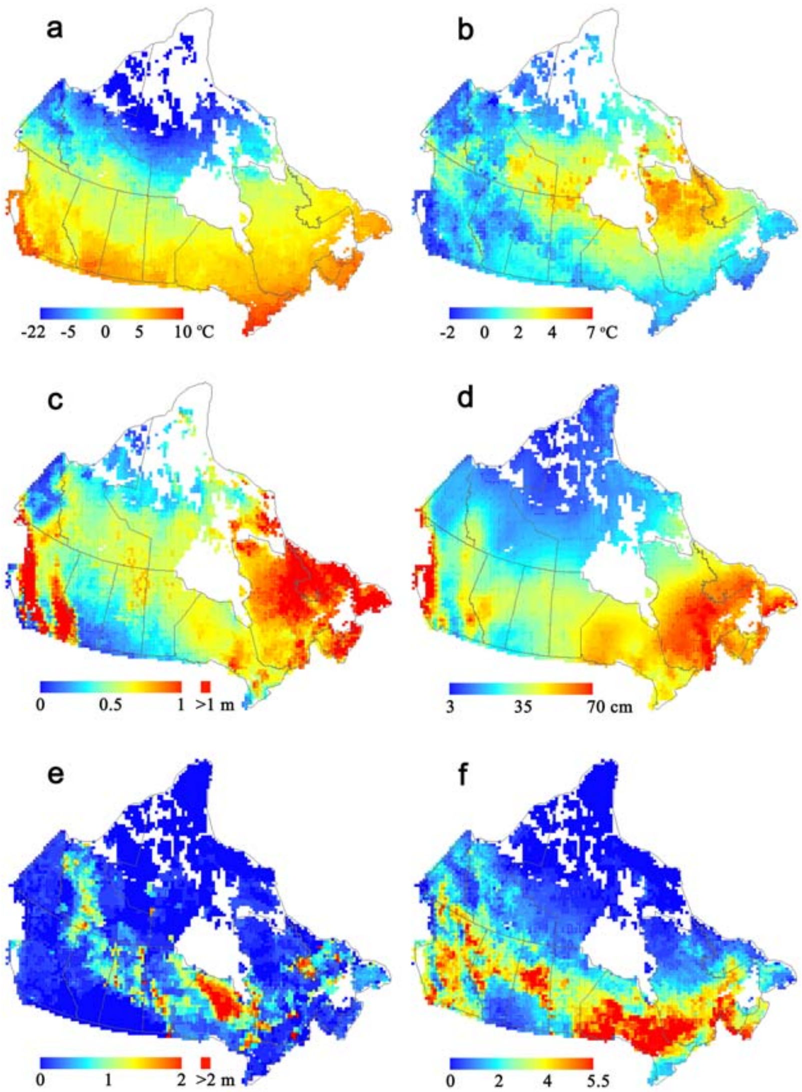

Figure 2. (a) simulated annual mean Ts at 20 cm depth, (b) the difference of annual mean Ts and Ta, (c) simulated average snow depth in winter (January, February and March), (d) total precipitation in summer (June, August, and September), (e) organic layer thickness, and (f) LAI in early August. The values in panels a - d were calculated as the averages over 1901-1995.

3. Results and Analysis

3.1. Tempo-Spatial Distributions of Soil Temperature during the Twentieth Century in Canada

3.1.1. Mean Soil Temperature

Figure 2a shows the simulated soil temperature (Ts) at 20 cm depth averaged over 1901-1995. The mean annual Ts ranged from –21.5 ˚C in the north to 9.5 ˚C in the south. This spatial distribution of Ts was similar to that of air temperature (Ta) ($R = 0.96$, $n = 6547$). *Smith and Burgess* [2000] compiled a database of ground temperature measurements for about 1000 sites in Canada. The near-surface ground temperature measurements were aggregated into 0.5-degree latitude/longitude grid cells using a mean and a range based on the measurements in a grid cell. Because the site conditions of the measurements could be very different from the averages of the grid cells we used in the modeling, the simulated Ts could be several degrees different from that of the measurements for some sites. The general pattern of Ts, however, agreed well with the measurements ($R = 0.93$, $n = 279$) (Figure 3).

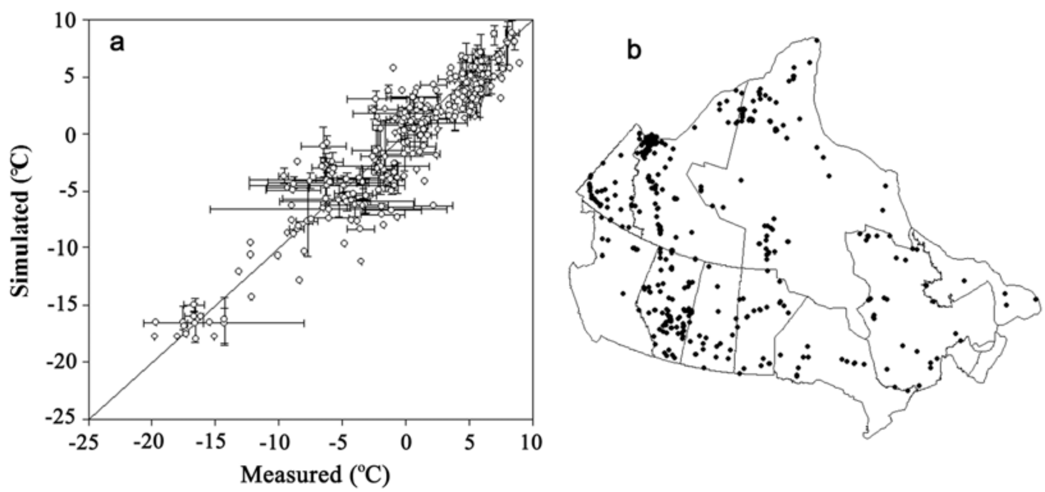

Figure 3. (a) comparison between simulated and measured Ts, and (b) the locations of the measurements. The measurements are from the ground temperature database of Canada [*Smith and Burgess,* 2000]. Observations were conducted at different depths and time periods. We aggregated the measurements into a mean and a range if more than one measurement was available in a grid cell. The range of the simulated Ts was the range of annual mean Ts for the corresponding observation period.

The Ts and Ta were different in both seasonal means and annual means because of the combination effects of snow, vegetation, organic layers and precipitation. On average for all the grid cells in Canada, the mean Ts in winter (January, February, and March) was 10.5 °C higher than that of the Ta, mainly due to snow insulation, which reduces heat loss from soil to the atmosphere. Thus, the spatial distribution of the Ts-Ta difference in winter was closely correlated with the snow depth ($R = 0.59$, $n = 6547$). In summer (July, August, and September), the mean Ts was 3.8 °C lower than that of the Ta averaged for all of Canada, mainly because of the combination effects of organic layers ($R = -0.55$, $n = 6547$), vegetation (LAI. $R = -0.38$, $n = 6547$), and precipitation ($R = -0.18$, $n = 6547$). Organic layers (including forest floor and peat layers) lower Ts because of their high porosity and usually high water content (such as wetlands), which increases evapotranspiration and heat capacity. The upper part of the forest floor is usually dry in summer and thus insulates the soil from warming [*Smith,* 1975]. The plant canopy shades the surface and enhances transpiration, therefore there is less energy available for warming the soil. Higher precipitation also corresponds to higher evapotranspiration and heat capacity.

For the annual means, the difference between Ts and Ta ranged from –2 to 7 ºC from place to place (Figure 2b), depending on the combined effects of snow cover, precipitation, organic layers, and vegetation (Figure 2c-f). Northeastern Canada had a higher Ts-Ta difference because of the thicker snow cover and lower LAI. The Ts-Ta difference was small or negative in western Canada because of the thick organic layer, shallow snow cover and high LAI. In the western coastal area, heavy rainfall and vegetation shading results in lower Ts, and the Ts-Ta difference was negative (Figure 2b). For the whole of Canada, snow insulation was the dominant factor determining the spatial distribution of the Ts-Ta difference ($R = 0.57$, $n = 6547$). The average difference between the annual means of Ts and Ta for the whole country was 2.5 ºC.

3.1.2 Changes of Soil Temperature during the Twentieth Century

Figure 4 compares the temporal variations of Ts with Ta during the twentieth century averaged for all the grid cells in Canada. The temporal pattern of the annual mean Ts (at 20 cm depth) was similar to that of Ta ($R = 0.84$, $n = 96$). The spatial distributions of the changes in Ts and Ta (Figure 5a, b) between the decades of 1986-1995 and 1901-1910 were also closely related (R = 0.78, n = 6549). This suggests that Ts generally responded to the forcing of Ta

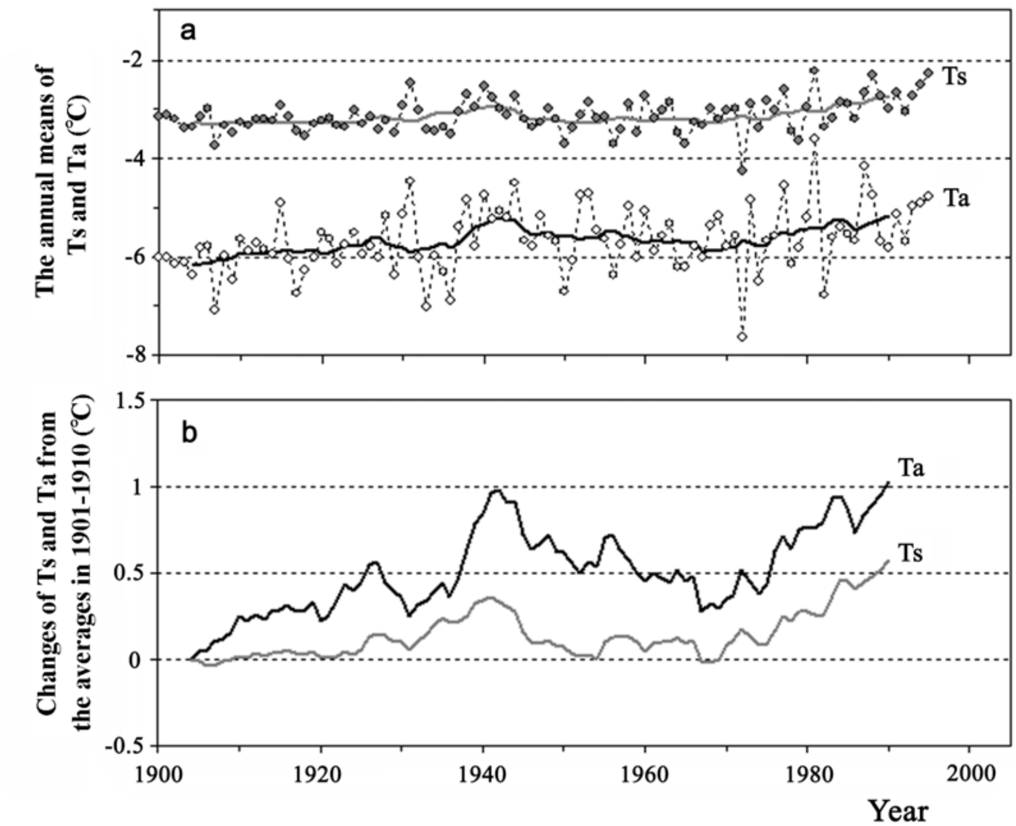

Figure 4. Changes of Ts in Canada comparing with the changes of Ta during the twentieth century. (a) the annual mean soil and air temperatures and their 10-year running averages, and (b) the changes of Ts and Ta from the decade 1901-1910 (10-year running averages minus the averages during 1901-1910).

The response of Ts to changes in Ta, however, is complex. The fluctuation of Ts was smaller than that of Ta, and the increase of Ts was smaller than that of Ta (Figure 4). From 1901 to 1995, Ta increased on average by 1.0 ˚C for the whole of Canada (based on the 10-year running average), while Ts increased by 0.6 ˚C. Ta increased in most of western Canada, especially in mid western areas, but decreased in eastern regions and around Hudson Bay (Figure 5a). The changes in Ts generally followed this spatial pattern, but with more spatial heterogeneity (Figure 5b). Changes in Ts and Ta during the twentieth century differed by -3 to 3 ˚C from place to place (Figure 5c). The increase in

Ts in mid western Canada was smaller than that of Ta, and the decrease in Ts in eastern Canada and around Hudson Bay was smaller than that of Ta. In the southwestern Northwest Territories, the magnitude of soil warming was larger than that of air, but on Baffin Island, the soil cooled while Ta did not change much. Figure 6 shows the temporal changes of Ts and Ta in these four regions during the twentieth century.

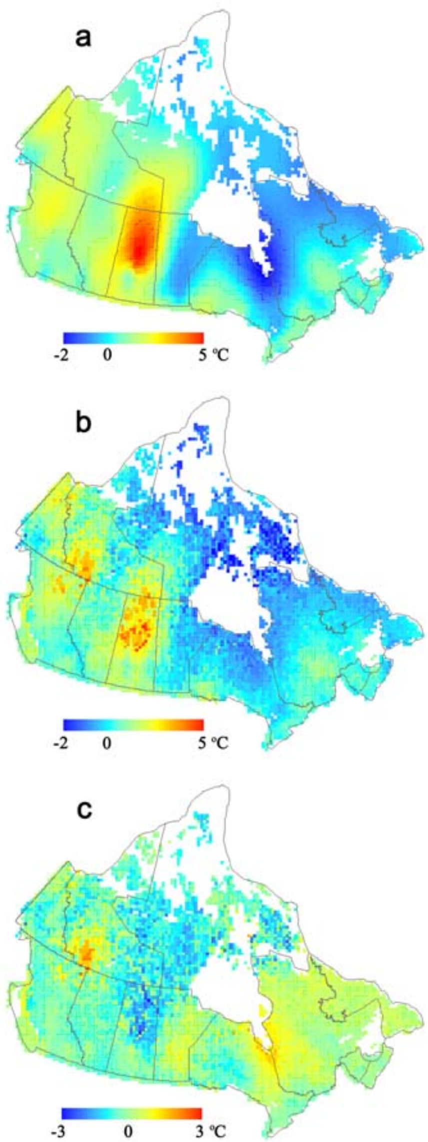

Figure 5. (a) changes in the annual mean Ta during the twentieth century, (b) changes in the annual mean Ts at 20 cm depth during the twentieth century, and (c) the difference between the changes of Ts and Ta during the twentieth century. The changes in Ts and Ta were calculated as the difference between the averages in 1986-1995 and in 1901-1910.

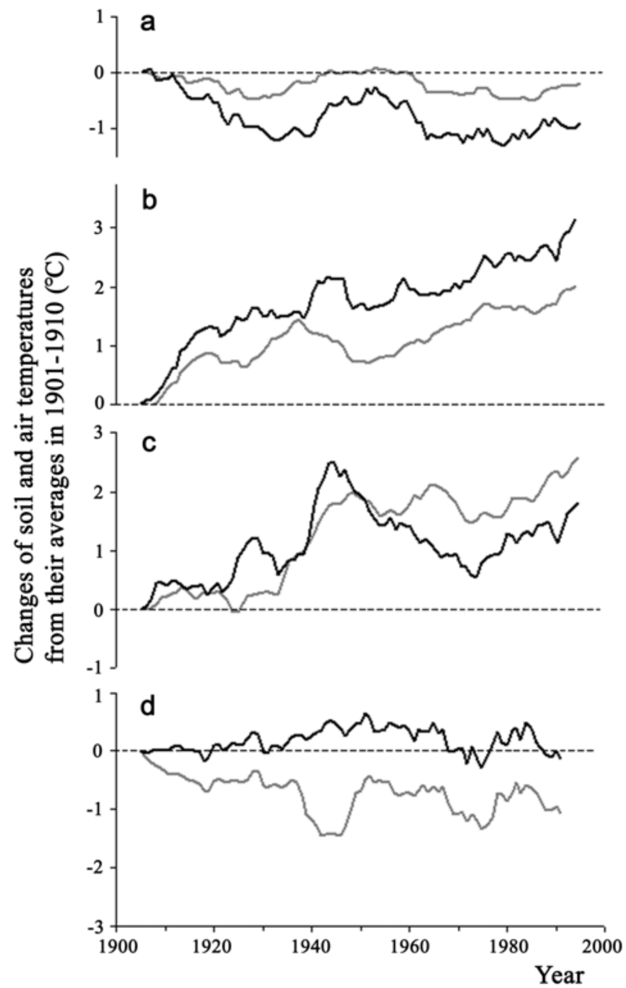

Figure 6. Comparison between the changes of Ts (grey curves) and Ta (black curves) in four regions. (a) southeast Canada (latitude: 47-55 °N, longitude: 77-83 °W), (b) mid western Canada (latitude: 55-65 °N, longitude: 100-110 °W), (c) southwest of Northwest Territories (latitude: 60-65 °N, longitude: 120-125 °W), and (d) Baffin Island (latitude: 63-68 °N, longitude: 65-75 °W). They were calculated as 10-year running averages minus the averages in 1901-1910.

The seasonal means of Ts and Ta also show different patterns. The increase in Ta occurred mainly in January to May, but the increase in Ts was small during these months; the increase in Ts was slightly higher than that of Ta in June and July; the changes in Ts and Ta were similar in August and September; and during October-December, Ta fluctuated from year to year while Ts was relatively stable (Figure 7). These patterns suggest that the difference in the rates of change between annual mean soil and air temperatures (Figure 4b) were mainly due to the slow response of Ts during January-May, with some local exceptions.

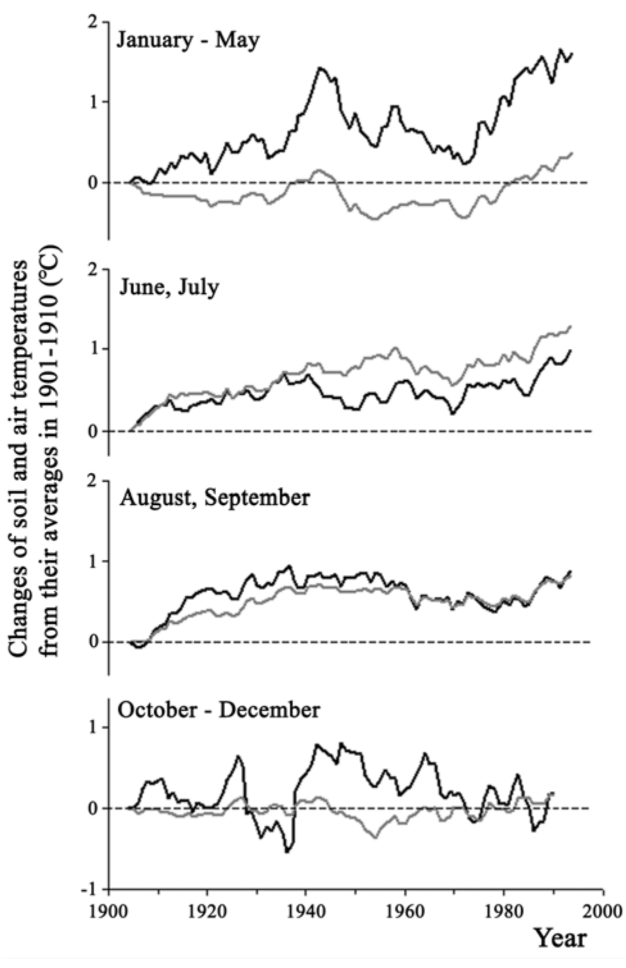

Figure 7. The seasonal distributions of the changes in Ts (grey curves) and Ta (black curves) averaged for the whole of Canada. They were calculated as 10-year running averages minus the averages in 1901-1910.

3.1.3. Observed Evidence about the Differentiation between Changes in Ts and Ta

The difference between the mean annual air and soil temperatures at high latitudes has long been recognized. This simulation also shows a large-scale, systematic difference between the change rates of air and soil temperatures. We checked some related measurements to see if they support this result. There are 14 climate stations in Canada with more than 20 years of soil temperature measurements, most of which are located in southern Canada [*Environment Canada,* 1994]. To compare the trends in Ts and Ta, we calculated the 5-year running averages for both Ts (at 20 cm depth) and Ta, and then calculated the deviation of these moving averages from the averages of the first 5-year period when both Ts and Ta were available (Figure 8). The results show that the increase in Ts was smaller than that of Ta at 8 stations (Figure 8a-h), the changes were similar at 4 stations (Figure 8i-l), and the increase in Ts was higher than that of Ta at 2 stations (Figure 8m, n). The differences between the changes in Ts and Ta were up to ±1.5 ˚C in about 20 years. These differences were comparable to the magnitudes of the changes in Ta. For example, Ta at station "a" (55.20 ˚N, 119.40 ˚W) increased 1.9 ˚C from 1966 to 1993 while the Ts increased only 0.5 ˚C (based on the 5-year running averages). At station "m" (58.38 ˚N, 116.03 ˚W), Ta increased 0.7 ˚C from 1960 to 1980 while Ts increased 3.2 ˚C. The station data were too sparse to validate the simulated spatial distributions, but analysis of observed data shows that the temporal variations in Ts and Ta can be significantly different. This result is consistent with the analysis of *Beltrami and Kellman* [2003a] for 10 climate stations in Canada, most of which have been included here.

Beltrami et al. [2003b] reconstructed ground surface temperature histories from temperature-depth measurements at 246 sites in mid and southern Canada (a region with latitudes less than 65 °N for western Canada, and less than 58 °N for mid and eastern Canada). The result shows that the ground surface warmed in southern and western Canada during the twentieth century. *Beltrami et al.* [2003b] estimated that the ground surface has warmed about 0.7 °C in the last 100 years on average for the study region. This magnitude is comparable to our simulated changes in soil temperature for this region (0.8 °C), and it was less than the change in Ta during the twentieth century (1.1 °C).

Recent observations in other regions also indicate differentiations between changes in Ts and Ta. For example, *Zhang T. et al.* [2001] analyzed a long-term soil temperature dataset (1889-1995) at a climate station in Russia (52 °N, 104 °E). The measurements show that both the magnitudes and patterns of changes between soil and air temperatures are different. Changes in precipitation (rainfall in summer and snowfall in winter) could be an important factor modifying the long-term patterns of the annual means and the seasonal means of Ts. *Harris et al.* [2003] found that early winter snow cover rather than mean air temperature is the dominant factor affecting the near surface permafrost temperature measured over 1987-2002 in the Swiss Alps. Similar results have been reported for some Russia climate stations and for a site in North Dakota, USA [*Pavlov,* 1994; *Schmidt et al.,* 2001], and about half of the ground temperature increase measured recently (1983-1988) on the Northern Slope of Alaska can be attributed to the changes in snow cover [*Stieglitz et al.*, 2003].

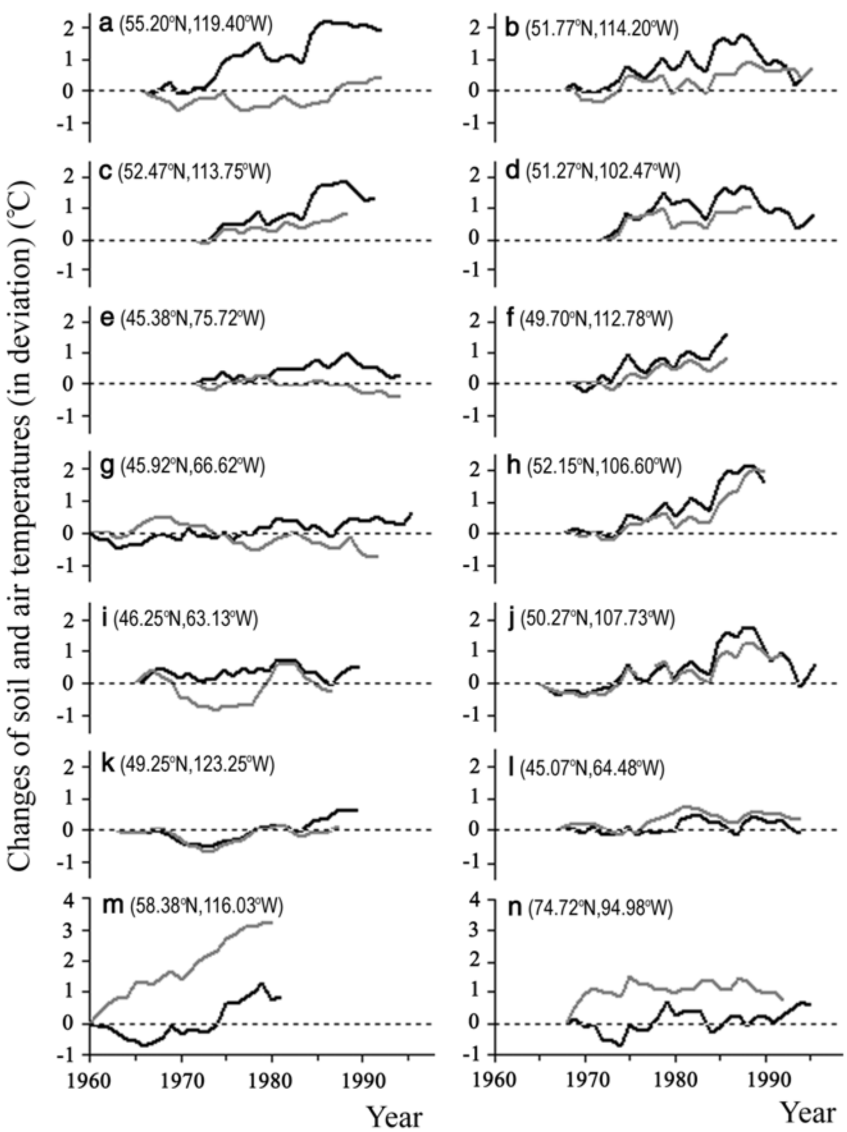

Figure 8. Comparison between changes in Ts (grey curves) and Ta (black curves) measured at climate stations in Canada. The measurements were from *Environment Canada* [1994]. The changes were calculated as 5-year moving averages minus the averages of the first 5-year period when both Ts and Ta data were available.

3.2. Mechanisms Which May Differentiate Changes in Soil and Air Temperatures

3.2.1. The Ta-snow-Ts Mechanism

The dynamics of snow, such as snowfall, snow depth, snow density, and the duration of snow cover on the ground, are very sensitive to Ta [*Kongoli and Bland,* 2000]. Therefore, changes in Ta would alter the duration, thickness, and density of snow cover, resulting in changes in the surface energy balance and heat conduction to the soil and therefore alter Ts. We refer to this process as the Ta-snow-Ts mechanism. To test this mechanism, we ran the NEST model for 1000 years with increasing and decreasing Ta by 0.03 ºC year^{-1} for 500 years from the current average conditions (corresponding to Ta increasing and decreasing 15 ºC, respectively) at 5 sites of different latitudes (45, 50, 55, 60 and 65 ºN). The longitudes were 76 ºW for the first site (Ottawa) and 105 ºW for the other 4 sites. The temperature change was distributed uniformly for all the months. The vapor pressure was changed with Ta by setting relative humidity as 60%. The amount and distribution of precipitation were kept as the baseline conditions. We assumed there were no vegetation and organic layers during the simulation.

Figure 9 compares the simulated Ts with the changes in Ta at site 1 (45 ºN, 76 ºW) and site 3 (55 ºN, 105 ºW). Mean Ts in winter was higher than Ta, especially when the Ta was low. At site 1, winter Ts decreased only about 3 ºC while Ta decreased from 0 ºC to -25 ºC. During the same period, the snow depth (the depth at the end of January) increased from 0 m to 1.6 m, and the number of days with snow depth greater than 0.01 m increased from 12 to 340 days. The insulating effect of snow cover generally increases logarithmically with its thickness [*Smith and Riseborough,* 2002; *Smith,* 1975]. Decreasing Ta caused an increase in snow depth, and the increased insulation effects of snow cover offset some of the cooling effects of Ta on Ts. At site 3, when the Ta in winter decreased from –4 ºC to –33 ºC, the snow depth increased from 0.1 m to 1.2 m, and the duration of snow cover increased from 83 to 343 days. The snow cover was thinner at this site because the precipitation was less (the annual precipitation was 500 mm at site 3, and 1010 mm at site 1). With decreasing Ta, Ts decreased faster at site 3 than at site 1, because the snow cover was thinner and its increase rate was slower than at site 1. At site 3, winter Ts decreased faster than Ta when Ta decreased more than 12 ºC, perhaps because the annual snow cover persisted for so long that it limited soil heat gain during the summer months, consistently reducing summer Ts below Ta in these years (Figure 9h).

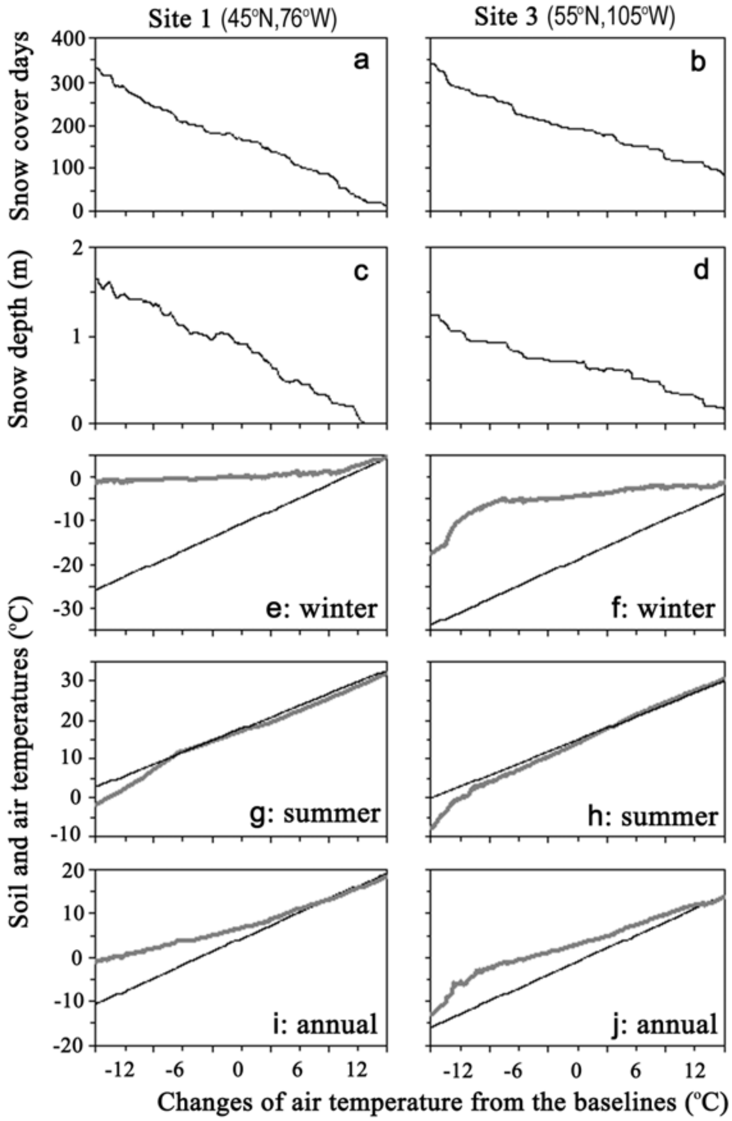

Figure 9. Changes in annual snow cover days (a, b), snow depth at the end of January (c, d), and the responses of Ts (grey curves) to Ta (black curves) in summer, winter and annual averages (e - j) simulated by changing Ta at site 1 and site 3. The snow depth at the end of January becomes 0 cm although snow cover exists in parts of the month (a, c).

The response of annual mean Ts to changes in Ta also shows a non-linear feature although the annual Ts-Ta difference was much smaller than in wintertime. At site 1, the annual mean Ts decreased from 13 ºC to -1 ºC while the annual mean Ta decreased from 13 ºC to -10 ºC; at site 3, the annual mean Ts decreased from 14 ºC to -3 ºC while the annual mean Ta decreased from 14 ºC to -11 ºC. These results indicate that the changes in

Ts and Ta were different due to the direct effects of changes in Ta on snow dynamics, which modulated the response of Ts to Ta.

This phenomenon could also be explained based on the fact that the annual mean Ts at high latitudes is 1 - 11 ºC higher than the annual mean Ta because of snow insulation effects [*Smith and Riseborough,* 2002; *Smith,* 1975], and Ts is close to Ta under snow-free conditions. Increases in Ta would eventually reduce snowfall and snow accumulation, and therefore would reduce the difference between the Ts and Ta induced by snow insulation. Thus, the increases of Ts would be 1 - 11 ºC less than the increases of Ta when Ta increases to such a level that there is no snow accumulation on the ground.

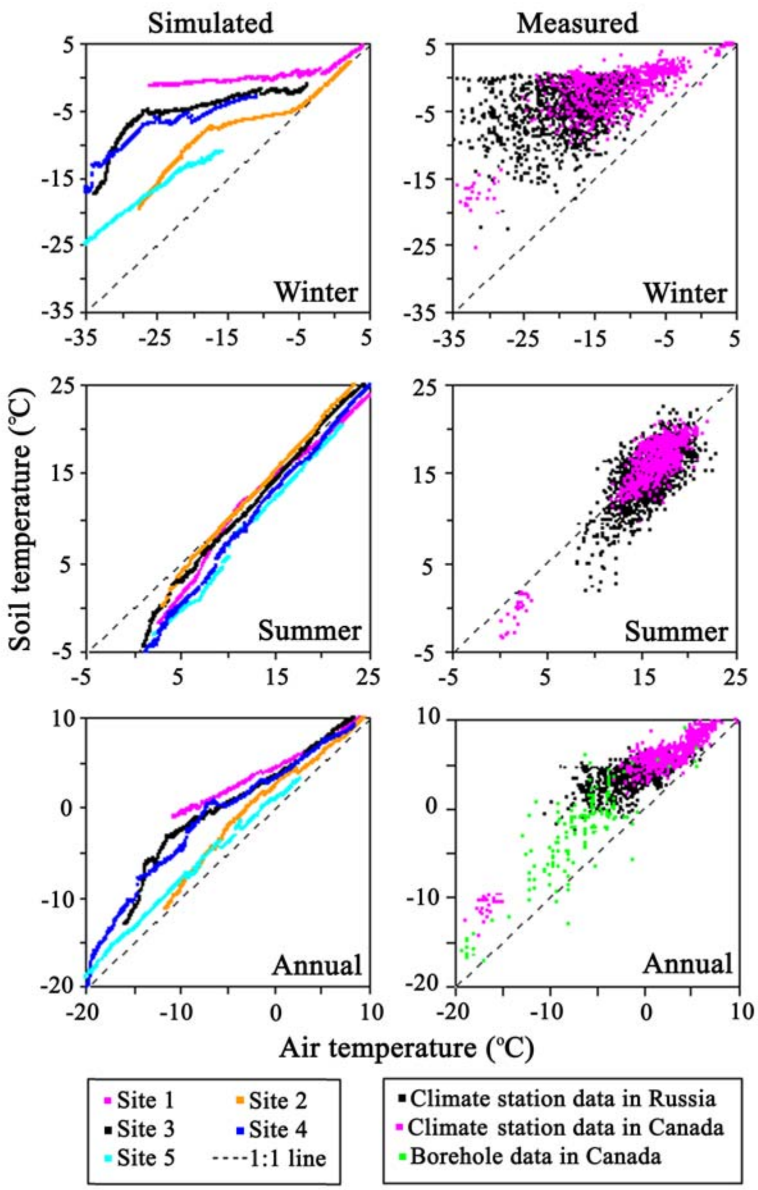

Figure 10. Simulated responses of Ts to changes in Ta at 5 sites, comparing with the scatter graphs of Ts and Ta measured in Canada and Russia. The climate station data in Canada were from *Environment Canada* [1994]; the borehole data in Canada (annual means of ground and air temperature) were from *Smith and Burgess* [2000]; and Russia climate station data were from the National Snow and Ice Data Center, University of Colorado, Boulder, USA.

This mechanism seems understandable, but is difficult to detect directly from measurements because precipitation and Ta fluctuate from year to year. Ts is measured at many locations around the world, but there are few long-term datasets. Scatter plots of annual mean Ts vs. annual mean Ta measured in Canada and Russia are compared to that of the model results (Figure 10). Common features are present in the scatter plots: the Ts in winter was higher than the Ta with wide scattering due to the variability in snow conditions; the Ts was close to the Ta during summertime; and the annual mean Ts responded to the Ta in a humped-shape with a large difference between the Ts and Ta when the annual mean Ta was from -15 ºC to 0 ºC, peaking where Ts was close to 0 ºC. The difference between the Ts and Ta also varied with location, mainly due to the difference in precipitation. The Ta-snow-Ts effects were smaller at sites 2 and 5, where precipitation was low (the annual precipitation was 390 mm and 230 mm, respectively) compared to site 1, where precipitation was high (annual precipitation of 1010 mm). The similarities between the simulated and measured results provide some supporting evidence about the effects of the Ta-snow-Ts mechanism, but this comparison should not be read as validation of the simulated results because the situations were different.

3.2.2. Seasonal Difference in Changes of Ta

In the twentieth century, increases in air temperature in Canada mainly occurred in winter and spring [*Serreze et al.,* 2000; *Zhang X. et al.,* 2000]. To test the significance of the seasonal distribution of changes in Ta in determining the changes in Ts (referred as the seasonal difference mechanism), we ran the NEST model using site 1 and site 3 as the baselines and increasing the annual mean Ta by 0.03 ºC year^{-1} with different seasonal distributions: increasing Ta by 0.03 ºC year^{-1} uniformly for all months, increasing Ta by 0.06 ºC year^{-1} for winter and spring months (December, and January – May), and increasing Ta by 0.06 ºC year^{-1} for summer and autumn months (June-November). Precipitation and its seasonal distribution were kept the same as the baseline conditions. Table 2 shows the simulated changes in the annual mean Ts at 20 cm depth in the year 100 (the annual mean Ta increased 3 ºC comparing to the baselines). The increase in Ts was smaller than the increase in Ta when there was snow on the ground (mainly in December, and January to April). In the summer-autumn warming treatment, the increase of Ts was close to the increase of Ta during the snow-free periods, with the soil remaining warmer in winter although Ta did not increase in winter months. The mean annual Ts increased 1.7 ºC in the winter-spring warming treatment for both sites 1 and 3, which was smaller than in the uniform warming treatment (2.0 ºC for these two sites) and in the summer-autumn warming treatment (2.6 ºC and 3.0 ºC for these two sites, respectively).

Table 2. Changes in snow depth and Ts in response to different seasonal distributions in Ta change.

Ta change.

Site	Site 1 (45 ºN, 76 ºW)		Site 3 (55 ºN, 105 ºW)	
	Snow[a](cm)	ΔTs[b] (ºC)	Snow[a](cm)	ΔTs[b] (ºC)
Baseline	119	-	73	-
Uniform warming	88	2.0	62	2.0
Winter-spring warming	76	1.7	63	1.7
Summer-autumn warming	93	2.6	64	3.0

[a] Mean snow depth in February.
[b] Difference of the mean annual Ts at 20 cm depth between the year 100 and year 0

To test if the lag in Ts will eventually reduce the difference between the rates of change in Ts and Ta, we ran the model for another 100 years after the above simulation but kept Ta as in the year 100. Ts at 35 m depth continued to increase for about 25 years before stabilized, but Ts in the upper layers (less than 1 m) increased by less than 0.1 ºC, suggesting that the Ts in surface layers was in equilibrium with the atmospheric climatic conditions. When the Ts at 35 m depth was in equilibrium with the atmospheric climatic conditions (Ts at this depth increased 0.5 ºC and 0.3 ºC from year 100 to 200 for sites 1 and 3, respectively), the difference between the Ts at 35 m depth and the Ta was similar to the difference between the Ts at upper layers and Ta (A small difference, about 0.1-0.2

ºC, was due to the thermal offset effect in upper layers and geothermal gradient in deeper layers). These results show that the whole soil thermal regime had a damped response to the winter-spring warming treatment.

Studies show that the insulating effect of snow on annual mean Ts is higher when the snow cover is thicker and the annual amplitude of Ta is larger [*Lachenbruch*, 1959; *Goodrich,* 1982]. In the above numerical experiments, the annual amplitude of Ta decreased in the winter-spring warming treatment, but increased in the summer-autumn warming treatment. These differences would alter the response of Ts to changes in Ta. All three treatments resulted in reduced snow depth, especially in the winter-spring warming treatment (Table 2), thereby reducing the insulating effect of the snow cover, leading to a lower increase in Ts. The uniform warming and summer-autumn warming treatments also enhanced evapotranspiration, which would contribute to the decreased response of Ts to Ta as well, especially at site 1, where precipitation was high.

3.2.3. Changes in Precipitation

Because snow cover and soil moisture can strongly modulate Ts through insulation and modifying evapotranspiration rate and soil thermal properties, changes in precipitation would directly alter snow conditions and soil moisture, thereby changing Ts and hence the relationship between Ts and Ta (referred as the precipitation mechanism). To test this mechanism, we ran NEST to equilibrium conditions for the above mentioned 5 sites using different precipitation and compared with the baseline conditions. We varied the amount of precipitation by ±10%, ±20% and ±50% without changing the distribution of precipitation in a year. Increasing precipitation increased Ts in winter but decreased Ts in summer, because of the corresponding increases in snow depth and evapotranspiration (Figure 11). The changes in the annual mean Ts could be different in direction and magnitude due to the opposite effects of snow and evapotranspiration on Ts, especially when the changes in precipitation were small (i.e., less than 20%). These numerical experiments show that the magnitudes of snow effects on Ts were larger than that of evapotranspiration, and Ts at higher latitudes was more sensitive to changes in precipitation. Due to the opposite effects of snow and evapotranspiration on Ts, the seasonal distribution of changes in precipitation would have different effects on Ts. For example, precipitation decrease in summer and/or increase in winter would increase Ts, but precipitation increase in summer and/or decrease in winter would decrease Ts.

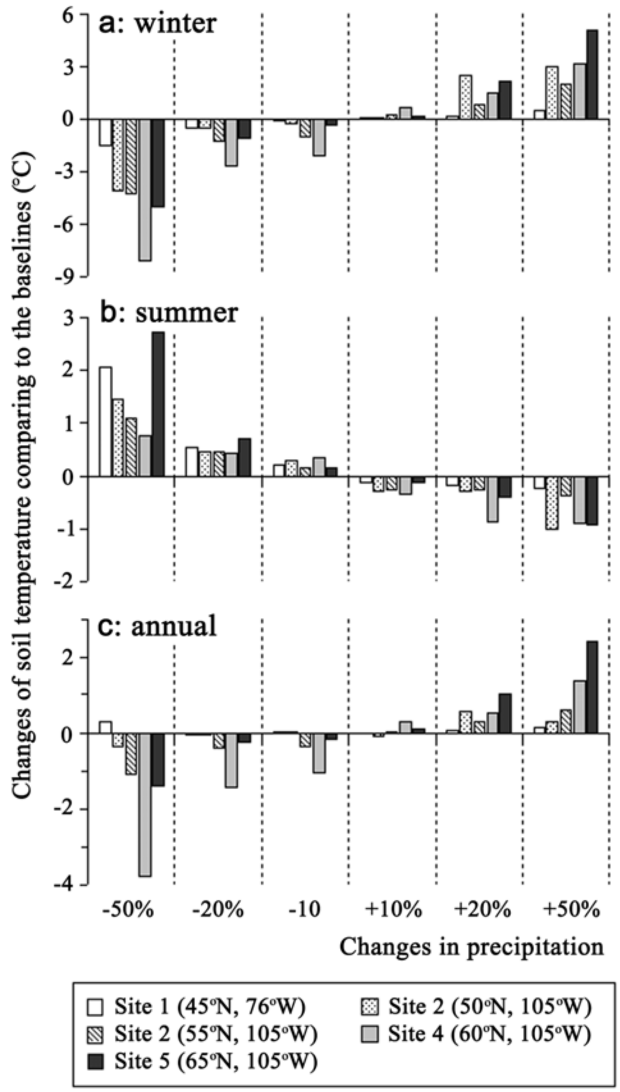

Figure 11. The effects of changes in precipitation on Ts. The amount of precipitation were varied in percentage without change its distribution. Notice the vertical scales are different among these panels.

3.2.4. Major Reasons for the Differences between the Changes in Ts and Ta in Canada during the Twentieth Century

The three mechanisms discussed above can help us to explain the differences between the changes in Ts and Ta in Canada during the twentieth century. The increase of Ta in mid western Canada (Figure 5a) would result in a smaller increase of Ts than that of Ta through the Ta-snow-Ts mechanism. Ta decreased in eastern Canada and around Hudson Bay (Figure 5a). The decrease in Ts would be smaller than that of Ta according to the Ta-snow-Ts mechanism. Figure 12d also shows a decrease in snow depth in the mid west and some increase in eastern Canada, which may be partly related to the Ta-snow-Ts mechanism. Changes in Ta (warming in the west and cooling in the east) were larger in winter and spring than in summer (Figure 12a, b). This seasonal difference would further buffer the response of Ts to the changes of Ta through the seasonal difference mechanism. Precipitation increased in southwest Northwest Territories with a corresponding increase in snow depth, resulting in a higher Ts. Decreases in precipitation on Baffin Island, however, resulted in a shallower snow depth and lower Ts (the precipitation mechanism. Figure 12c, d).

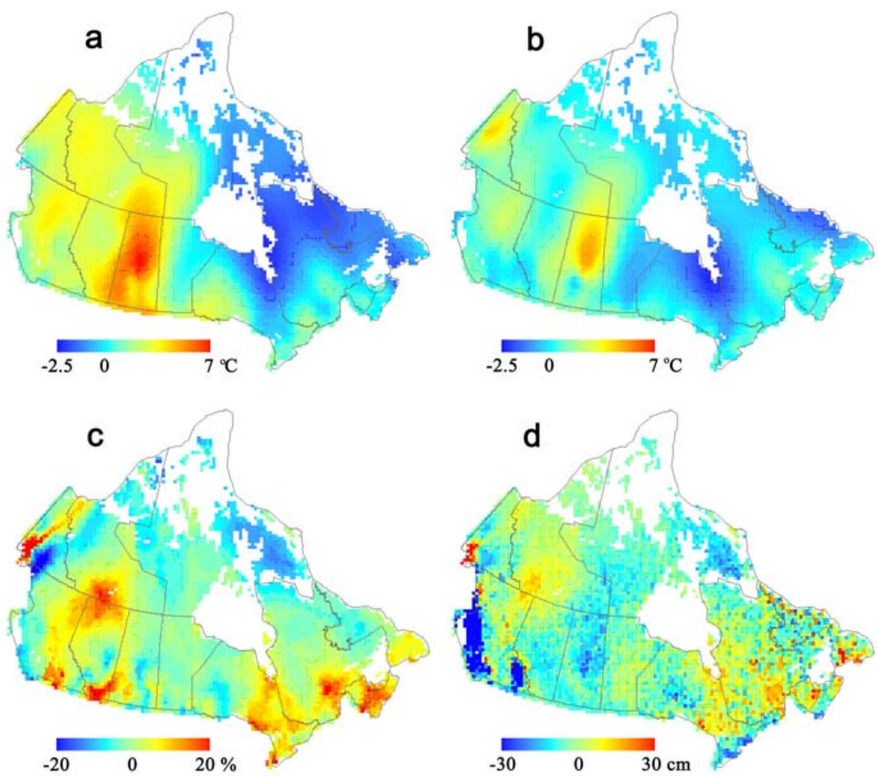

Figure 12. Changes in (a) Ta in winter (January, February and March), and (b) Ta in summer (June, July and August), (c) annual precipitation, and (d) snow depth from the decade 1901-1910 to the decade 1986-1995.

These three mechanisms reflect the physical interactions among atmospheric climate, snow dynamics and soil thermal regime, and their effects on soil temperature are different in tempo-spatial patterns and in time scales. The precipitation mechanism is significant, especially at higher latitudes, and precipitation changes dramatically with space and time. The Ta-snow-Ts mechanism can buffer the long-term response of Ts to atmospheric climate change at high latitudes, especially when precipitation is high. The seasonal difference mechanism is important for most of the high latitudes because changes in Ta mainly occur in winter and spring [*Serreze et al.,* 2000; *Zhang X. et al.,* 2000; *McCarthy et al.,* 2001]. Because of the combined effect of these mechanisms, the future changes in Ts could differ from that of Ta as projected by GCMs, and the difference could significantly modify the impacts of atmospheric climate change on permafrost distribution, as well as ecological and biogeochemical processes. These results also confirm that the historical ground surface temperature derived from geothermal data cannot be directly interpreted as changes in atmospheric temperature [*Pavlov,* 1994; *Schmidt et al.,* 2001; *Zhang T. et al,* 2002; *Beltrami,* 2002; *Harris et al.,* 2003],

4. Conclusions

The tempo-spatial distributions of soil temperature in Canada during the twentieth century were simulated using a process-based model. The results show complex responses of soil temperature to changes in atmospheric climate. The major conclusions can be summarized as the following:

- The general spatial distribution of the mean annual Ts was similar to that of Ta, but the mean annual Ts differed from the mean annual Ta by –2 to 7 ˚C from place to place, due to the combined effects of snow cover, precipitation, soil organic layers, and vegetation. The average difference averaged for the whole of Canada was 2.5 ˚C.
- Ts responded to the forcing of Ta in a complex way. The annual mean Ts increased 0.6 ˚C in the twentieth century averaged for the whole Canada, which was smaller than the increase of the annual mean Ta (1.0 ˚C). The difference between the changes of Ts and Ta mainly occurred in winter and spring.
- The differences between the changes in Ts and Ta in Canada during the twentieth century can largely be explained by three mechanisms: the Ta-snow-Ts mechanism, seasonal differences in changes of Ta, and changes in precipitation.

This study shows that Ts and Ta at high latitudes could be different in their mean values and in their long-term patterns, and it is difficult to directly infer Ts from Ta, or vice versa. *Groisman et al.* [1994] suggested that the strong spring warming in recent decades may be because of the retreat of spring snow cover in the Northern Hemisphere. Our study shows that changes in Ta and precipitation and their seasonal distributions all could differentiate the response of Ts to Ta, mainly because of the associated changes in snow cover. Therefore it is essential to couple atmospheric climate, snow, soil thermal and hydrological dynamics, as well as ecological and biogeochemical processes in predicting and assessing the impacts of climate change.

We tried to use the best available datasets in this long-term, national-scale, spatial modeling, but some weaknesses of the input data used for this study need to be identified. First, there are relatively few climate stations in the high latitudes (particularly north of 60°N), and generally data quality is poorer before the 1940s and in mountainous regions, especially for precipitation. Secondly, it is only in recent decades that large-scale, high-resolution vegetation data have become available through remote sensing techniques. In this long-term study, long-term changes in vegetation and soil organic matter were not considered. Future work would couple NEST with biogeochemical models (i.e. the InTEC model. *W. Chen et al.*, 2000a; 2000b), which not only can simulate ecological processes with the impacts of soil thermal and hydrological dynamics, and but also can improve the simulation of soil temperature, especially the effects of disturbance (forest fire and harvesting).

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