Global Change Biology (2007) **13**, 2425–2438, doi: 10.1111/j.1365-2486.2007.01450.x

Energy feedbacks of northern high-latitude ecosystems to the climate system due to reduced snow cover during 20th century warming

E. S. EUSKIRCHEN*, A. D. McGUIRE† and F. S. CHAPIN III*

*Institute of Arctic Biology, University of Alaska Fairbanks, Fairbanks, AK 99775, USA, †US Geological Survey, Alaska Cooperative Fish and Wildlife Unit, University of Alaska Fairbanks, Fairbanks, AK 99775, USA

Abstract

The warming associated with changes in snow cover in northern high-latitude terrestrial regions represents an important energy feedback to the climate system. Here, we simulate snow cover-climate feedbacks (i.e. changes in snow cover on atmospheric heating) across the Pan-arctic over two distinct warming periods during the 20th century, 1910–1940 and 1970–2000. We offer evidence that increases in snow cover–climate feedbacks during 1970–2000 were nearly three times larger than during 1910–1940 because the recent snow-cover change occurred in spring, when radiation load is highest, rather than in autumn. Based on linear regression analysis, we also detected a greater sensitivity of snow cover–climate feedbacks to temperature trends during the more recent time period. Pan-arctic vegetation types differed substantially in snow cover–climate feedbacks. Those with a high seasonal contrast in albedo, such as tundra, showed much larger changes in atmospheric heating than did those with a low seasonal contrast in albedo, such as forests, even if the changes in snow-cover duration were similar across the vegetation types. These changes in energy exchange warrant careful consideration in studies of climate change, particularly with respect to associated shifts in vegetation between forests, grasslands, and tundra.

Keywords: albedo, climate change, Pan-arctic, snow melt, snow return

Received 24 October 2006; revised version received 30 May 2007 and accepted 17 May 2007

Introduction

Variations in seasonal snow cover and the associated changes in albedo and atmospheric heating are key components of the climate system. These variations are considered valuable indicators of climate change because of their sensitivity to temperature (Karl *et al*., 1993; Groisman *et al*., 1994), and studies have documented a retreat of snow in response to high-latitude warming in recent decades (Brown, 2000; Dye, 2002). This retreat of snow is important due to the resulting feedback: as snow retreats, less solar energy is reflected to space, and more energy is absorbed and transferred to the atmosphere (Groisman *et al*., 1994), causing a positive snow/albedo feedback loop that reinforces warming. These variations are particularly influential in high-latitude (above 50°N) terrestrial regions, where seasonal contrasts in albedo and surface heating may be as large within vegetation types as they are among vegetation types (Betts & Ball, 1997; Eugster *et al*., 2000).

We may gain a better understanding of the effect of warming on snow cover, and in turn, the effect of decreases in snow cover on warming, by examining changes in snow cover under warming periods of differing magnitude. Although the temperature record during the 20th century shows a large degree of variability, it is marked by two distinct warming periods: (1) 1910–1940 and (2) 1970–2000 (Fig. 1). In the earlier period, the trends in warming are not as strong as those in the later period, but in the earlier period there is a greater range in the air temperature trends across latitudinal bands. Between 1910 and 1940, trends in air temperature increase, based on the slopes of least squares linear regression, ranged from $0.0078^{\circ}\,\mathrm{yr}^{-1}$ in the region between 50 and 55°N to $0.0383^{\circ}\,\mathrm{yr}^{-1}$ between 70 and 75°N. From 1970 to 2000, trends in air temperature increase were more uniform, ranging from

Correspondence: E. S. Euskirchen, tel. +1 907 474 1958, fax +1 907 474 6967, e-mail: ffese@uaf.edu

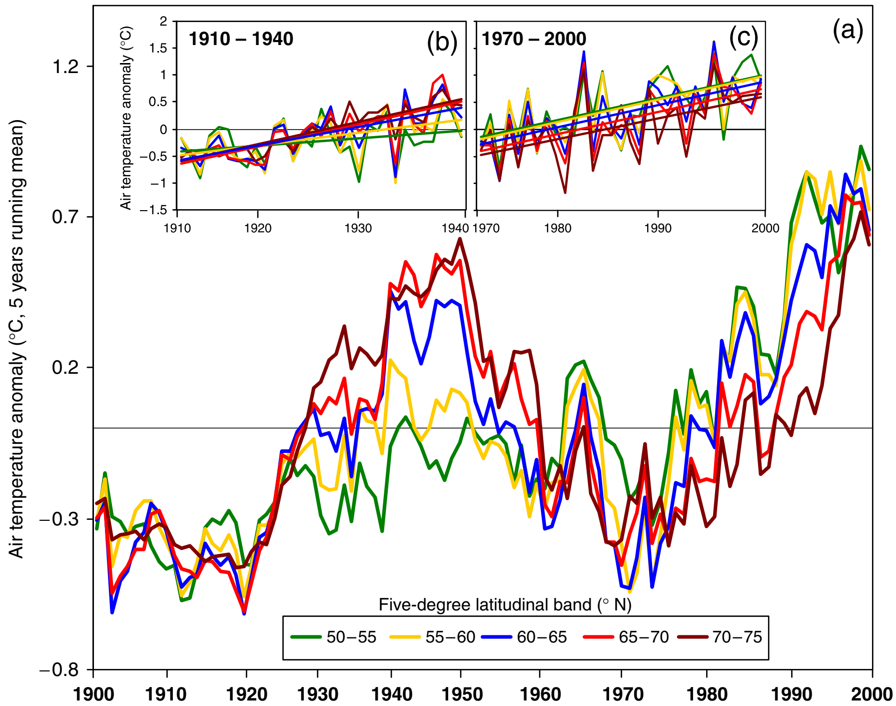

Fig. 1 Changes in terrestrial land surface air temperature by 5° latitudinal bands based on data from the Climate Research Unit (CRU, Mitchell *et al.*, 2004). In (a), 5-year moving averages of the temperature anomaly from 1901 to 2000 are presented. In (b), trends in air temperature for 1910–1940, based on the slopes of least squares linear regression, ranged from 0.0078° yr^{-1} in the region between 50 and 55°N to 0.0383° yr^{-1} between 70 and 75°N. In (c), trends in air temperature for 1970–2000 were more uniform, ranging from approximately 0.0377° yr^{-1} between 50 and 55°N to 0.0353° yr^{-1} between 70 and 75°N.

approximately 0.0377° yr^{-1} between 50 and 55°N to 0.0353° yr^{-1} between 70 and 75°N. These two periods of warming thus provide an interesting contrast against which to evaluate how changes in snow cover due to increases in temperature impact atmospheric heating.

In a previous study (Euskirchen *et al.*, 2006), we used a water balance model (Vörösmarty *et al.*, 1989) to estimate changes in snow cover, and then validated these estimates against a 29-year time series (1972–2000) of remotely sensed satellite data (Dye, 2002), examining the timing of snow melt in the spring, the return of snow in the fall, and the duration of the snow-free season. Both datasets indicated that from 1972 to 2000, there were overall decreases in the duration of the snow-free period, with trends toward earlier melt in the spring and later return of snow in the fall. Our estimates of reductions in snow cover in the high latitudes during 20th century warming agree with other studies (Foster *et al.*, 1992; Brown, 2000; Stone *et al.*, 2002; Rikiishi *et al.*, 2004; Chapin *et al.*, 2005).

Here, we build upon our previous analyses of changes in snow cover and examine patterns in snow melt, snow return, and the duration of the snow-free season as they impact atmospheric heating. We perform our analyses for the 1910–1940 and 1970–2000 time periods over the arctic-boreal land area above 50°N, a region of approximately $3.4 \times 10^6\,km^2$. We examine these patterns across several different categories, including the entire region between 50 and 75°N, latitudinal bands between 50 and 75°N, and dominant vegetation types (Fig. 2). Finally, we consider these estimates of atmospheric heating due to changes in snow-cover duration with respect to feedbacks in atmospheric heating that may be caused by changes in carbon storage in the soils and vegetation of terrestrial high-latitude regions.

Methodology

Overview

We performed simulations of snow cover and incoming solar irradiance (R_S) between 1910–1940 and 1970–2000 at half-degree latitude by longitude spatial resolution with the TERRESTRIAL ECOSYSTEM MODEL (TEM, version 5.1, Euskirchen *et al.*, 2006). We then computed changes in snow cover for these two time periods, examining changes in both snow melt in the spring and snow

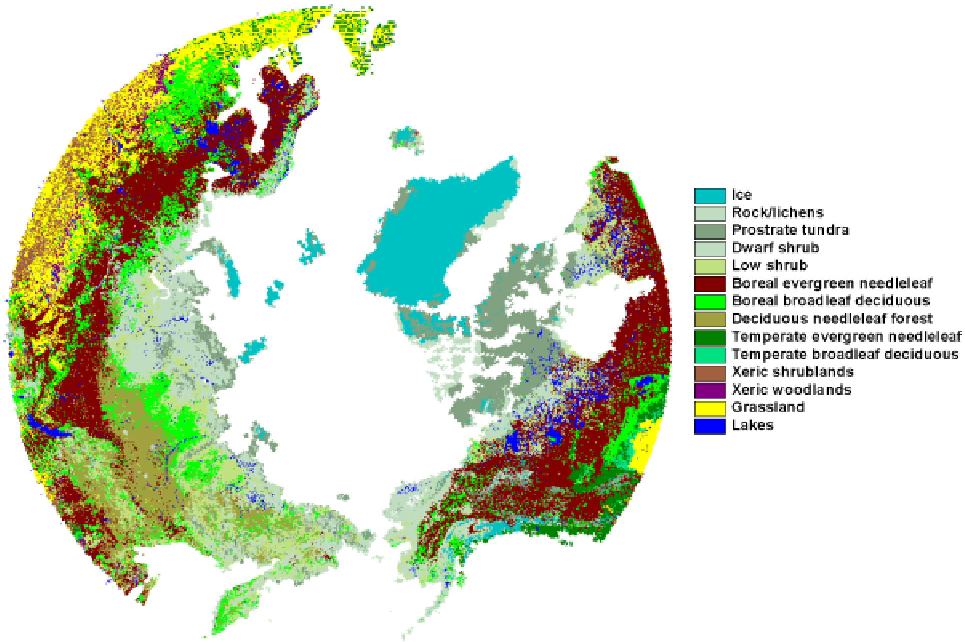

Fig. 2 Cover types of the terrestrial land area north of 50°N, based on continental and regional-scale delineations of the arctic and subarctic. The percentage of the land area occupied by each cover types is: ice (7.7%); rock/lichens (2.2%); prostrate shrub tundra (8.2%); dwarf shrub tundra (11.6%); low shrub tundra (5.4%); boreal evergreen needleleaf (24.1%); boreal broadleaf deciduous (11.6%); deciduous needleleaf (11.3%); temperate needleleaf (2.8%); temperate broadleaf (0.3%); xeric shrubland (1.5%); xeric woodland (0.9%); grassland (9.7%); and lakes (2.7%).

return in the fall. We used the estimates of R_S in combination with literature-derived energy budget parameters pertaining to net radiation and heat flux over snow-free and snow-covered ecosystems to estimate the effect of changes in snow cover on atmospheric heating. We performed our analyses over the land area north of 50°N, with an emphasis on trends in dominant vegetation types and in the latitudinal bands between 50 and 75°N.

Snow and net solar irradiance in the TEM

We used the water balance model (Vörösmarty *et al.*, 1989) within the TEM (version 5.1) to estimate changes in snow cover. TEM is a process-based, global-scale ecosystem model that incorporates spatially explicit (half-degree latitude by longitude) data pertaining to vegetation (Fig. 2), climate, soil, and elevation to estimate monthly pools and fluxes of carbon and nitrogen in the terrestrial biosphere (Raich *et al.*, 1991). The water balance model produces monthly gridded estimates of snow pack. Snow pack accumulates whenever mean monthly temperature is below −1 °C, and snow melt occurs at or above −1 °C. At elevations of 500 m or less, the model removes the entire snow pack, plus any new snow by the end of the first month with temperatures above −1 °C. At elevations above 500 m, the melting process requires 2 months above −1 °C, with half of the first month's snow pack retained to melt during the second month (Vörösmarty *et al.*, 1989). To estimate the date of snow melt (or snow return) from the monthly estimates of snow pack, we incorporated an algorithm that uses a 'ramp' between monthly temperatures. Linear interpolations of data for monthly air temperature and the month(s) preceding snow melt (or snow return), the month of snow melt (or snow return), and the month following snow melt (or snow return) are performed. For example, to calculate the date of snow melt when all snow has disappeared by April, approximately 30 points are interpolated between mean monthly March and April air temperature to determine the 15 points for the first half of March, and approximately 30 points are interpolated between mean monthly air temperature in April and May to determine the 15 points for the second half of April. The length of the snow-free season is calculated by subtracting the date of snow return by the date of snow melt for each half-degree latitude by longitude grid cell.

Parameterization of surface energy balance and estimates of atmospheric heating

To determine changes in surface energy balance between snow-covered and snow-free ground in the

Table 1 Mean values of energy budget parameters across dominant vegetation types north of 50°N

Vegetation type	Location and reference	Energy budget parameter	Pre-snow-melt	Post-snow-melt	Pre-snow-return	Post-snow-return
Low shrub tundra	Rouse *et al.* (2003)	R_n (% of R_S)	16	52	40	8
		Atmospheric heating (% of R_n)	23	96	56	96
	68°N, 133°W	$MJ\,m^{-2}\,day^{-1}$	0.4	8.7	2.8	0.7
Dwarf and prostrate tundra	Rouse *et al.* (2003)	R_n (% of R_S)	14	52	35	5
		Atmospheric heating (% of R_n)	31	84	56	89
	68°N, 133°W	$MJ\,m^{-2}\,day^{-1}$	0.15	7.6	2.8	0.4
Boreal evergreen	Liu *et al.* (2005)	R_n (% of R_S)	42	80	29	12
		Atmospheric heating (% of R_n)	98	77	78	78
	63°N, 145°W	$MJ\,m^{-2}\,day^{-1}$	11.4	15.5	3.2	0.9
Boreal deciduous broadleaf	Liu *et al.* (2005)	R_n (% of R_S)	56	62	23	5
		Atmospheric heating (% of R_n)	47	96	66	80
	63°N, 145°W	$MJ\,m^{-2}\,day^{-1}$	5.4	9.9	2.16	0.4
Boreal deciduous conifer	Ohta *et al.* (2001)	R_n (% of R_S)	19	51	22	10
		Atmospheric heating (% of R_n)	82	95	89	49
	62°N, 129°E	$MJ\,m^{-2}\,day^{-1}$	1.5	7.5	2.8	0.5
Grassland/shrub	Liu *et al.* (2005)	R_n (% of R_S)	18	29	21	7
		Atmospheric heating (% of R_n)	28	90	66	100
	63°N, 145°W	$MJ\,m^{-2}\,day^{-1}$	0.7	6.6	2.0	0.7
Temperate broadleaf deciduous	Moore *et al.* (1996)	R_n (% of R_S)	89	90	51	22
		Atmospheric heating (% of R_n)	91	96	53	97
	42°N, 78°W	$MJ\,m^{-2}\,day^{-1}$	19.0	24.1	4.7	2.6
Temperate evergreen needleleaf	Restrepo & Arain (2005)	R_n (% of R_S)	57	83	27	14
		Atmospheric heating (% of R_n)	91	98	96	91
	42°N, 80°W	$MJ\,m^{-2}\,day^{-1}$	9.5	13.0	4.0	1.7

'Heating' refers to atmospheric heating, given both as a percentage of net radiation (R_n), calculated based on measured heat flux to the atmosphere [sensible (H) plus latent (LE) fluxes], and in $MJ\,m^{-2}\,day^{-1}$, calculated as $R_S \times (R_n/R_S) \times (H + LE)/R_n$. Values of incoming shortwave radiation (R_S) are calculated in TEM. Estimated values of R_n, H, and LE are obtained from the references, as indicated with the corresponding latitude and longitude of the study sites. The study of Moore *et al.* (1996) reported 3 years of data and that of Rouse *et al.* (2003) reported 2 years of data. For these two studies, we averaged values across all reported years of data. All other studies were based on data collected over an annual cycle.

TEM, TERRESTRIAL ECOSYSTEM MODEL.

dominant high-latitude ecosystems, we first compiled empirical data from the peer-reviewed literature on the seasonal (with 'seasonal' referring to the pre-snow-melt, post-snow-melt, pre-snow-return, and post-snow-return time periods) fluxes of sensible heat (H), latent heat (LE), and net radiation (R_n). Because energy flux data were not available over a full seasonal cycle for the xeric shrubs and xeric woodlands in snow-covered regions, and these vegetation types comprised a small percentage of the terrestrial land area (0.9–1.5%; Fig. 2), we assigned them values for the grasslands and deciduous forests, respectively. All of the studies employed a combination of radiometers and eddy-covariance measurement systems to estimate these fluxes. We used mean seasonal values across all reported measurement years for a given vegetation type (Table 1).

We then calculated seasonal values of incoming solar irradiance (R_S) in TEM for each half-degree grid cell. These estimates are based on the methodology of Turton (1986) and are attenuated by the input cloud cover with a correction based on that described in Chang (1968). Following methodology similar to that of Chapin *et al*. (2005), we estimated seasonal atmospheric heating for each vegetation type in each half-degree grid cell by multiplying R_S by the proportion of incoming R_S that is absorbed by the land surface (R_n/R_S) times the proportion of R_n that is transferred to the atmosphere ($\{H + LE\}/R_n$). We then compared pre- and post-snow-melt and pre- and postreturn energy budgets to estimate the changes in snow melt and snow return on atmospheric heating: {[Daily atmospheric heating post-snow-melt (*or pre-snow-return*)]–[Daily atmospheric heating pre-snow melt (*or post-snow-return*)] × [Change in snow-cover duration]}, where the daily atmospheric heating is in units of $\mathrm{MJ\,m^{-2}\,day^{-1}}$ and the change in snow-cover duration is in $\mathrm{days\,yr^{-1}}$. The estimates of heating are then averaged over the length of the snow free season, as calculated in the water balance model. Estimates of heating are then presented at the decadal time scale in $\mathrm{W\,m^{-2}}$ based on the mean annual changes in the surface energy.

Model input and simulations

Our input climate datasets consisted of monthly cloudiness (%), precipitation (mm), and air temperature (°C) data for the years 1901–2000 obtained from the Climate Research Unit (CRU TS 2.0) database (Mitchell *et al*., 2004). The gridded input elevation map (used in the water balance model) is based on 10 min digital global elevation data (NCAR/Navy, 1984). The input vegetation map (Fig. 2) includes fractional cover types for each half-degree grid cell for the terrestrial land area north of 50°N. This map was created in order to better delineate arctic and subarctic vegetation types important in our study. Because most global cover type maps have highly aggregated vegetation types, we used continental or regional scale maps when possible and used a few of the global maps for ice and lake cover. The Alaska region is based on Fleming (2000), which had one 'tundra' category for all of interior and southern Alaska. The 'tundra' category was replaced in all parts of the state, with the exception of the North Slope and the Seward Peninsula, which incorporated the background vegetation from the USGS 1 km spatial resolution Land Cover dataset (Fleming, 2000). The North Slope (Muller *et al*., 1999) and Seward Peninsula (CAVM Team, 2003) coverages were generated from Landsat MSS data. Eurasian vegetation was based on Belov (1990). Data from the International Biosphere Program (Loveland *et al*., 2000) at a 1 km resolution was used to differentiate the 'woodland tundra' category in this region. Western Eurasian vegetation was based on Melillo *et al*. (1993). Canadian vegetation was based on the 1 km land cover vegetation dataset from the Canada Centre for Remote Sensing, which was prepared from multitemporal Advanced Very High Resolution Radiometer (AVHRR) data (Cihlar & Beaubien, 1998). Lakes were taken from Melillo *et al*. (1993), Cihlar & Beaubien (1998), and Loveland *et al*. (2000), and the ice category was also from Melillo *et al*. (1993).

We performed model simulations with transient climate data for the years 1901–2000 and analyzed the output data from the two warming periods, 1910–1940 and 1970–2000. To initialize the simulation, we ran TEM to equilibrium for all grid cells north of 50°N following the protocol of Zhuang *et al*. (2003), which consisted of using the mean climate from 1901 to 1930, as the equilibrium climate in 1900.

Results

Seasonal changes in atmospheric heating across vegetation types

In general, the eddy covariance data indicate that changes in atmospheric heating from snow-covered to snow-free ecosystems in the spring are larger than during the transition from snow-free to snow-covered ecosystems in the fall (Table 1). Across our study region, the greatest seasonal differences in atmospheric heating occurred during the spring in the tundra, where the increase in heating from the snow-covered to snow-free ground was 7–8 $\mathrm{MJ\,m^{-2}\,day^{-1}}$. Regions occupied by boreal and temperate evergreen and deciduous forests showed seasonal differences in heating between snow-covered and snow-free ground in the spring of 4–5 $\mathrm{MJ\,m^{-2}\,day^{-1}}$, while areas of grasslands/shrubs and deciduous needleleaf forests varied by ~6 $\mathrm{MJ\,m^{-2}\,day^{-1}}$. In the fall, the return of snow caused a more uniform reduction in heating across the vegetation types, approximately 1–2 $\mathrm{MJ\,m^{-2}\,day^{-1}}$ (Table 1).

Changes in snow cover

Decreases in snow-cover duration in the region between 50 and 75°N were larger from 1970 to 2000 (−0.22 $\mathrm{days\,yr^{-1}}$) compared with 1910–1940 (−0.16 $\mathrm{days\,yr^{-1}}$; Table 2; Fig. 3). This resulted from both later dates of snow return in this region from 1910 to 1940 (0.11 $\mathrm{days\,yr^{-1}}$) than from 1970 to 2000 (0.06 $\mathrm{days\,yr^{-1}}$) and earlier dates of melt from 1970 to 2000 (−0.16 $\mathrm{days\,yr^{-1}}$) than from 1910 to 1940 (−0.08 $\mathrm{days\,yr^{-1}}$). The mean length of the snow free season in the region between 50 and 75°N was 154 days between 1910 and 1940 and 157 days between 1970 and 2000. Among the vegetation types, the temperate broadleaf and deciduous forests displayed both the smallest (~ −0.10 $\mathrm{days\,yr^{-1}}$ between

Table 2 Mean change in snow cover (days yr^{-1}); with a negative value indicating an earlier trend (or shorter total duration for the 'total') and a positive value indicating a later trend as simulated with the TEM

Vegetation type or latitudinal band	Time period	Change in snow cover duration (days yr^{-1})			Mean length of snow-free season (days)
		Snow melt	Snow return	Total	
Low shrub tundra	1910–1940	−0.07	0.10	−0.17	95
	1970–2000	−0.14	0.08	−0.22	97
Dwarf and prostrate tundra	1910–1940	−0.09	0.10	−0.19	116
	1970–2000	−0.16	0.08	−0.24	118
Boreal evergreen	1910–1940	−0.02‡	0.12	−0.14	149
	1970–2000	−0.18	0.06	−0.24	152
Boreal deciduous broadleaf	1910–1940	−0.04	0.11	−0.15	140
	1970–2000	−0.18	0.04	−0.22	143
Boreal deciduous conifer	1910–1940	−0.08	0.08	−0.16	115
	1970–2000	−0.12	0.04	−0.16	117
Grassland/Xeric Shrub	1910–1940	+ 0.02‡	0.12	−0.10	184
	1970–2000	−0.20	0.00	−0.20	189
Temperate broadleaf deciduous	1910–1940	−0.03	0.06	−0.09	192
	1970–2000	−0.26	0.08	−0.34	198
Temperate evergreen needleleaf	1910–1940	+ 0.01‡	0.06	−0.05	171
	1970–2000	−0.34	0.04	−0.38	177
50–55°N	1910–1940	+ 0.01‡	0.06	−0.05	205
	1970–2000	−0.19	0.02	−0.21	210
55–60°N	1910–1940	−0.06	0.10	−0.16	187
	1970–2000	−0.21	0.02	−0.23	192
60–65°N	1910–1940	−0.08	0.13	−0.21	156
	1970–2000	−0.17	0.07	−0.24	159
65–70°N	1910–1940	−0.07	0.13	−0.20	126
	1970–2000	−0.14	0.11	−0.25	127
70–75°N	1910–1940	−0.09	0.11	−0.20	99
	1970–2000	−0.11	0.07	−0.18	99
50–75°N	1910–1940	−0.08	0.11	−0.16	154
	1970–2000	−0.16	0.06	−0.22	157

Values are based on least squares linear regression between the day of snow melt or snow return over the specified time period, with negative values of snow-melt change indicating earlier melt and positive values of snow return indicating later return. All changes, except those marked with (‡), are statistically significant ($P<0.0001$).

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1910 and 1940) and the largest (0.34–0.38 days yr^{-1} between 1970 and 2000) decreases in snow-cover duration. These forests are primarily located in the 50–60°N latitudes (Fig. 2) where trends in air temperature were smallest between 1910 and 1940 and greatest between 1970 and 2000. Across latitudinal bands (Table 2), there were generally greater ranges between the total change in snow-cover duration between 1910 and 1940 than between 1970 and 2000, corresponding to the greater differences among trends in air temperature across latitudinal bands between 1910 and 1940 than between 1970 and 2000 (Fig. 1). Changes in date of snow return were more geographically uniform from 1970 to 2000 than from 1910 to 1940 (Fig. 3). Later date of snow return was most pronounced in western Eurasia and least pronounced in central and eastern Canada from 1910 to 1940 (Fig. 3a). The changes in date of snow melt were due to a shift from trends toward later snow melt in western Eurasia and south-central North America between 1910 and 1940 (Fig. 3b) to trends toward earlier snow melt in these regions between 1970 and 2000 (Fig. 3d).

Changes in atmospheric heating due to changes in snow cover

Changes in atmospheric heating from 1970 to 2000 compared with 1910–1940 were magnified relative to changes in snow-cover duration because changes in the atmospheric heating of snow-covered and snow-free ground were greater in spring than in autumn (Table 1). For example, in the grassland/xeric shrub vegetation type, the decrease in snow-cover duration was 0.10

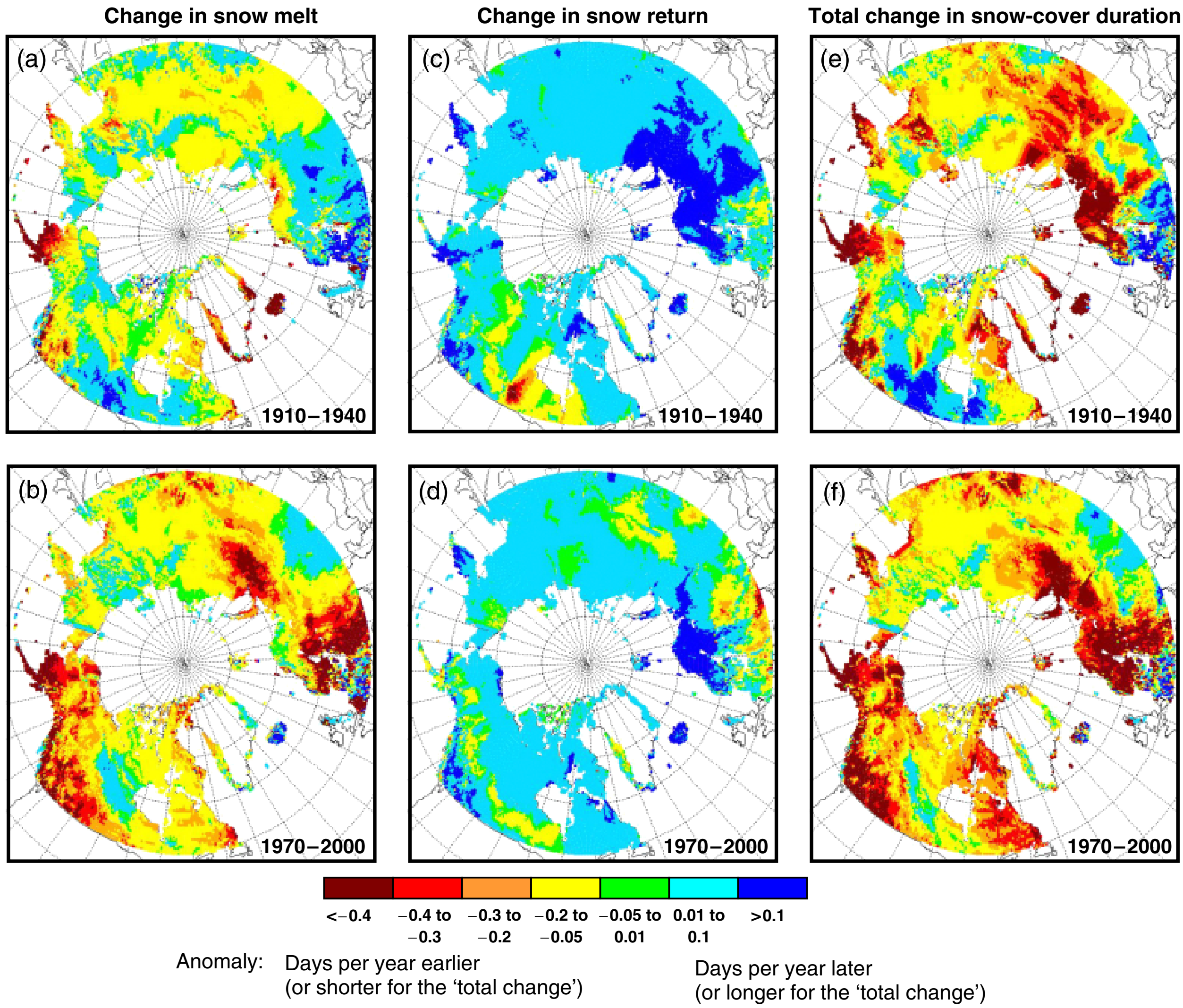

Fig. 3 Spatial distribution of changes in snow cover (days yr^{-1}) for changes in snow melt (a, b); changes in snow return (c, d) and changes in total length of snow-cover duration (e, f).

days yr^{-1} between 1910 and 1940 and 0.20 days yr^{-1} between 1970 and 2000 (Table 2), but there was a disproportional increase in heating between 1970 and 2000 (0.5 W m^{-2} $decade^{-1}$), when decreases in snow-cover duration occurred due to earlier snow melt compared with 1910–1940 (0.05 W m^{-2} $decade^{-1}$), when the decreases in snow-cover duration occurred primarily due to later snow-return (Fig. 4a). Likewise, in the latitudinal band between 70 and 75°N, the decrease in snow-cover duration was 0.20 days yr^{-1} (primarily due to later snow return) between 1910 and 1920 and 0.18 days yr^{-1} (primarily due to earlier melt) between 1970 and 2000 (Table 2), but the increase in heating in the later period (1.03 W m^{-2} $decade^{-1}$) was much larger than the increase in heating in the earlier period (0.25 W m^{-2} $decade^{-1}$, Fig. 4b).

Large differences in atmospheric heating were also noted between the vegetation types with large seasonal-albedo contrast (e.g. tundra) compared with low seasonal-albedo contrast (e.g. forests) even if they exhibited similar changes in snow-cover duration (Fig. 4a). For example, although the prostrate/dwarf shrub tundra and boreal forests displayed comparable decreases in snow-cover duration from 1970 to 2000 (approximately 0.23 days yr^{-1}, Table 2), the increase in heating was larger for tundra ($\sim$1.4 W m^{-2} $decade^{-1}$) than boreal forests (0.9 W m^{-2} $decade^{-1}$) due to the high contrast in albedo between snow-covered and snow-free ground in the tundra.

To assess the change in atmospheric heating from changes in snow cover (W m^{-2}, anomaly) per change in average annual temperature (°C, anomaly) we performed least squares linear regression analyses for

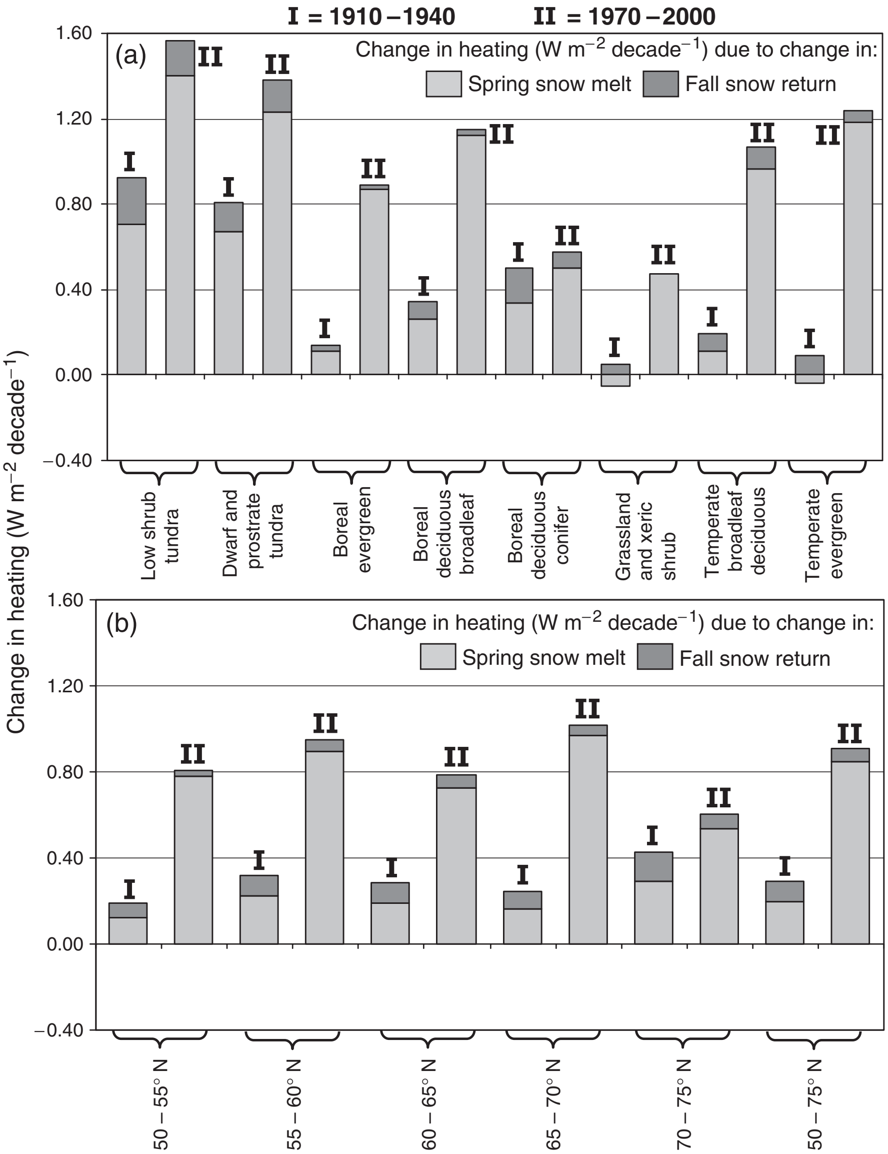

Fig. 4 Mean changes in energy (W m^{-2} decade^{-1}) due to changes in snow cover in the spring and autumn by vegetation type (a) and latitudinal bands, weighted by the proportion of each vegetation type in a given band (b) for the two time periods, 1910–1940 and 1970–2000. Positive energy change is equal to heating, and negative energy changes are equal to cooling.

snow return and snow melt across vegetation types and latitudinal bands for the two time periods (Table 3). Generally, changes in average air temperature had a greater effect on atmospheric heating from changes in spring snow melt, with the dry and wet tundra and temperate evergreen needleleaf forests showing the greatest sensitivity during both the 1910–1940 and 1970–2000 time periods (based on the significance of the *P*-values). Across the latitudinal bands, the greatest sensitivity from 1910 to 1940 occurred for snow melt in the 50–55°N range, while between 1970 and 2000, the greatest sensitivity generally occurred for snow melt at higher latitudes, especially between 60 and 70°N. Overall, these analyses illustrated a greater sensitivity to changes in air temperature during the 1970–2000 time period ($P < 0.005$) compared with the earlier 1910–1940 time period ($P < 0.05$, Table 3).

The anomaly maps in Fig. 5 show large spatial and interdecadal variability and an overall heating effect due to decreases in snow cover. There were essentially no areas where a cooling effect due to changes in total snow-cover duration was seen in 1970–2000 (Fig. 5f) that was not seen between 1910 and 1940 (Fig. 5c). In fact, most areas that showed a cooling effect from changes in snow-cover duration between 1910 and 1940 (Fig. 5c) switched to heating effect between 1970

Table 3 Change in atmospheric heating (W m^{-2}, anomaly) per change in average annual temperature (°C, anomaly)

Vegetation type	Years	Time of snow change	Change in anomaly (W m^{-2} per °C)	R^2	P-value
Dry tundra	1910–1940	Melt	0.84	0.56	<0.0001
		Return	0.14	0.36	0.0003
	1970–2000	Melt	0.94	0.58	<0.0001
		Return	0.14	0.49	<0.0001
Wet tundra	1910–1940	Melt	0.66	0.43	<0.0001
		Return	0.12	0.38	0.0002
	1970–2000	Melt	0.82	0.54	<0.0001
		Return	0.12	0.56	<0.0001
Boreal evergreen	1910–1940	Melt	0.83	0.34	0.0006
		Return	0.18	0.11	0.0637
	1970–2000	Melt	0.89	0.44	<0.0001
		Return	0.02	0.17	0.0215
Boreal deciduous broadleaf	1910–1940	Melt	0.79	0.33	0.008
		Return	0.08	0.19	0.0083
	1970–2000	Melt	0.99	0.46	<0.0001
		Return	0.05	0.18	0.0182
Boreal deciduous needleleaf	1910–1940	Melt	0.24	0.23	0.0075
		Return	0.15	0.35	0.0003
	1970–2000	Melt	0.29	0.33	0.0008
		Return	0.17	0.50	<0.0001
Grasslands/xeric shrubs	1910–1940	Melt	0.61	0.25	0.0038
		Return	0.05	0.06	0.1775
	1970–2000	Melt	0.76	0.43	<0.0001
		Return	0.07	0.17	0.0199
Temperate deciduous broadleaf	1910–1940	Melt	0.71	0.30	0.0015
		Return	0.18	0.08	0.1252
	1970–2000	Melt	0.83	0.35	0.0005
		Return	0.19	0.18	0.0159
Temperate evergreen needleleaf	1910–1940	Melt	1.24	0.53	<0.0001
		Return	0.33	0.43	<0.0001
	1970–2000	Melt	1.03	0.41	<0.0001
		Return	0.43	0.49	<0.0001
50–55°N	1910–1940	Melt	0.82	0.73	0.001
		Return	0.09	0.64	0.009
	1970–2000	Melt	0.90	0.81	<0.0001
		Return	0.09	0.72	0.0074
55–60°N	1910–1940	Melt	0.81	0.50	0.054
		Return	0.09	0.56	0.002
	1970–2000	Melt	0.90	0.63	0.0002
		Return	0.08	0.54	0.0043
60–65°N	1910–1940	Melt	0.51	0.46	0.0912
		Return	0.09	0.51	0.0554
	1970–2000	Melt	0.67	0.49	0.0160
		Return	0.09	0.51	0.0047
65–70°N	1910–1940	Melt	0.59	0.58	0.0003
		Return	0.12	0.62	0.0004
	1970–2000	Melt	0.76	0.61	<0.0001
		Return	0.11	0.54	0.0054
70–75°N	1910–1940	Melt	0.60	0.33	0.009
		Return	0.14	0.36	0.0004
	1970–2000	Melt	0.65	0.47	0.0002
		Return	0.13	0.42	0.0002

(*contd.*)

Table 3. (*Contd.*)

Vegetation type	Years	Time of snow change	Change in anomaly ($W\,m^{-2}$ per °C)	R^2	P-value
50–75°N	1910–1940	Melt	0.67	0.52	0.0311
		Return	0.11	0.54	0.0134
	1970–2000	Melt	0.78	0.60	0.0033
		Return	0.10	0.55	0.0044

Analyses are based on coefficients obtained from least squares linear regression for snow return and snowmelt across vegetation types and latitudinal bands (weighted by the proportion of each vegetation type in a given band) for the two time periods.

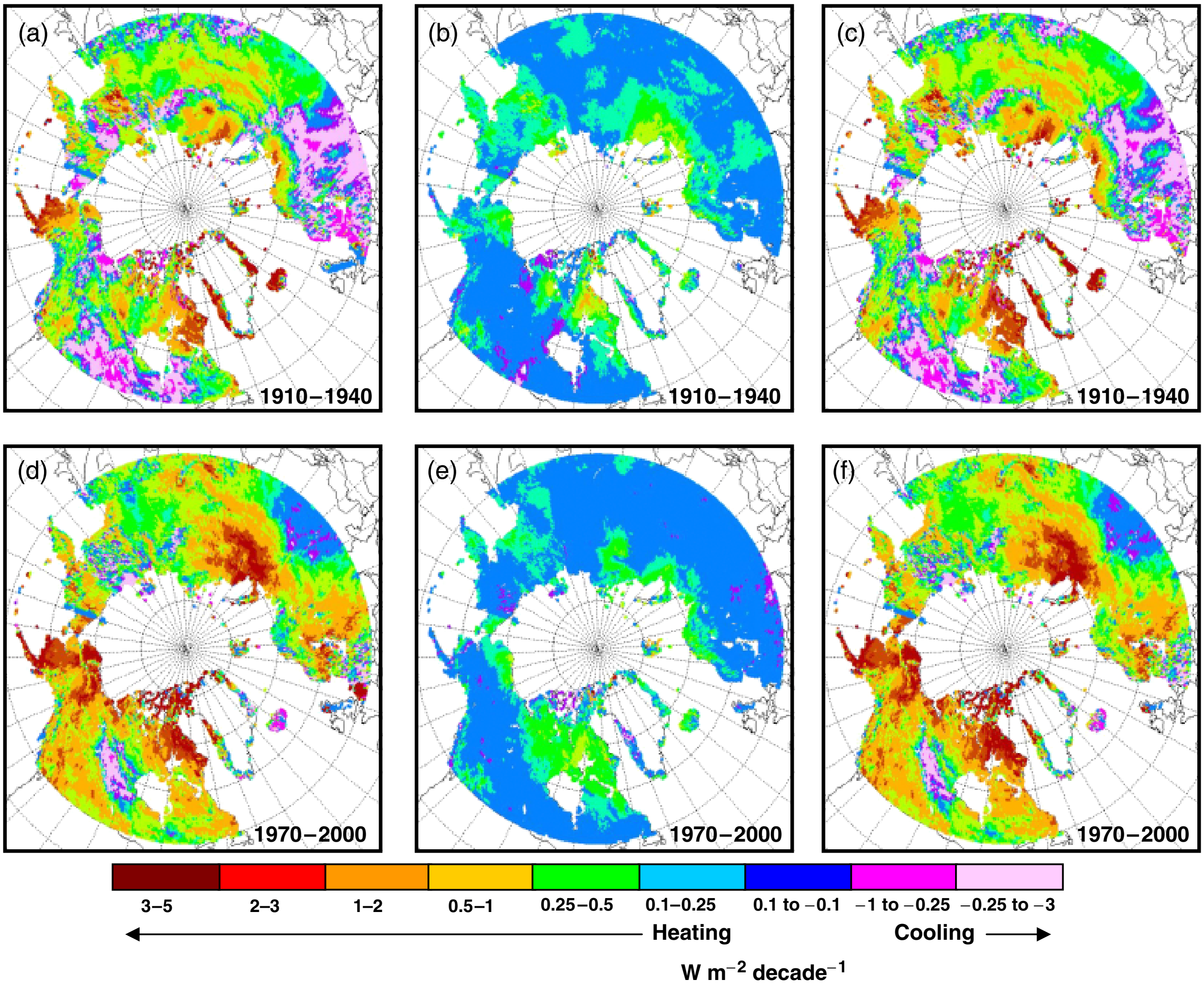

Fig. 5 Changes in energy ($W\,m^{-2}\,decade^{-1}$) due to changes in the timing of snow melt for 1910–1940 (a) and 1970 (b); changes in energy due to changes in the timing of snow return for 1910–1940 (c) and 1970–2000 (d); changes in energy due to changes in the length of snow-cover duration for 1910–1940 (e) and 1970–2000 (f). Each half-degree grid cell is area weighted by the proportion of each vegetation type in a given grid cell, based on the map in Fig. 2.

and 2000 (Fig. 5f). Across the entire study domain, our findings indicate that changes in energy due to changes in snow cover show a heating effect of +0.3 $W\,m^{-2}\,decade^{-1}$ during 1910–1940 with the magnitude increasing to +0.9 $W\,m^{-2}\,decade^{-1}$ between 1970 and 2000 (Figs 4b, 5c and 5f).

Discussion

This study evaluated snow cover-climate feedbacks (i.e. the effects of changes in snow cover on changes in atmospheric heating) during two periods of recent warming, 1910–1940 and 1970–2000. Our study supports the conclusion that decreases in snow cover during both of these periods increased atmospheric heating of the atmosphere, with the greater heating increases occurring in vegetation types with high contrast in albedo between snow-free and snow-covered ground, and in the later, 1970–2000, time period. In the discussion below, we examine the importance of considering the albedo contrasts between vegetation types, as well as the importance of examining changes in both snow melt and snow return in relation to changes in atmospheric heating. We also explore factors underlying the temperature increases during the two time periods as they relate to changes in snow cover, and how possible changes in vegetation structure may affect atmospheric heating. Finally, we discuss the relative influences of increases in heating due to changes in snow cover change vs. mitigation in heating due to increases in carbon sequestration by terrestrial ecosystems.

Model evaluation

Importance of considering the interactions between changes in snow cover and energy exchange of ecosystems. Although there is a long history of the inclusion of seasonal estimates of albedo in climate models (Viterbo & Betts, 1999), the concept that albedo may differ widely across vegetation types and snow-free and snow-covered ground received little attention until the past decade. Large-scale interdisciplinary field campaigns, such as the boreal ecosystem–atmosphere study (BOREAS), collected information pertaining to vegetation-specific levels of albedo over snow-free and snow-covered ground (Betts & Ball, 1997). This information was later included in climate models, greatly increasing their predicative capacities in snow-covered regions (e.g. Roesch *et al*., 1999; Viterbo & Betts, 1999). The approach presented in this study benefits from the incorporation of a relatively fine-scale representation of vegetation and data from eddy covariance studies. In our analyses, we included information from vegetation types with distinct patterns of surface energy exchange and considered the fractional cover of vegetation types within each half-degree grid cell. For example, we chose to explicitly define the deciduous needleleaf category because this vegetation type comprises approximately 11% of the terrestrial land area north of 50°N (Fig. 2), and its surface energy balance differs from that of boreal evergreen conifer or boreal deciduous broadleaf forests (Eugster *et al*., 2000). Although a large amount of eddy covariance data of carbon, water, and energy exchanges have become available from the network of FLUXNET sites worldwide (e.g. Baldocchi *et al*., 2001), these data have generally not been used in parameterizing global climate models even though they contain valuable information on surface energy budgets. We found that energy flux eddy covariance data reported in the peer-reviewed literature were useful in estimating snow cover–climate feedbacks only if they included at least one full annual cycle of energy flux measurements and the seasonal snow surface conditions. As further energy flux estimates from these eddy covariance studies become available, our approach permits us to readily incorporate them into our analyses.

Model validation. As mentioned in 'Introduction,' our estimates of trends in snow melt and snow return agreed well with the analyses of Dye (2002) based on satellite data. Pearson's rank-correlation coefficients between the water balance model estimates of snow melt, snow return, and the duration of the snow free period with those of the satellite data ranged between 0.36 and 0.73 (Euskirchen *et al*., 2006). The correlation of 0.36 occurred in extremely high latitudes (>70°N) where the instrumental climate data are scarce, and hence may influence the accuracy of the model input data. Discrepancies between modeled data and the Dye (2002) data might also reflect uncertainties associated with interpreting the remote sensing data. Factors that may reduce the reliability of remotely sensed snow-cover data include low solar illumination and high solar zenith angles and cloud cover (Dye, 2002). Nevertheless, the comparisons between the two datasets were strong enough that the trends in snow melt, snow return, and the duration of the snow-free season could not be attributable to an artifact of the methods. Moreover, our estimates of changes in snow cover agree well with those of other studies that find a recent shortening of the snow season in high latitudes by 0.15–0.35 days yr^{-1}, generally due to earlier snow melt (Stone *et al*., 2002; Chapin *et al*., 2005).

Consideration of snow-cover variation in terms of both snow melt and snow return

One striking result from this study was the difference in snow cover-climate feedback due to a change in spring snow melt vs. a change in snow return (Fig. 4). This is due to the asymmetry of snow melt and snow return over the region. That is, snow melt primarily occurs between the vernal equinox and the summer solstice while snow return generally occurs between the autumn equinox and winter solstice. The greater amount of solar input at the time of snow melt is primarily responsible for greater snow cover–climate feedback in the spring. During the dynamic spring shoulder season, solar radiation increases rapidly, the snow melts, net radiation is enhanced, ground thaw occurs, and the sensible and latent heat fluxes become substantial. A 1 °C increase in spring air temperature may increase surface energy fluxes by 7–10% of average values, a magnitude that is considerably greater than at any other time of year (Rouse *et al*., 2003). Before snow return in the fall, there is already little-absorbed solar radiation.

Consequently, if the snow pack is established later due to warmer temperatures, the increase in absorbed solar radiation would be small. Similarly, with an earlier creation of the snow pack, the decrease in absorbed solar radiation would be relatively minor. Thus, we saw a disproportional effect on atmospheric heating during 1970–2000 when changes in snow cover were primarily due to earlier melt than 1910–1940 when changes in snow cover were largely due to later snow return.

Time-period differences

The reasons underlying the increasing temperatures during the 1910–1940 and 1970–2000 in our study region are complex and not fully understood. The two factors thought to compose this warming are natural climate variability and anthropogenic forcings (McGuire *et al*., 2006). The trend in increasing temperatures as a result of natural climate variability in high latitudes is due to weather patterns that trigger warm air advection from lower latitudes, including the Arctic (North Atlantic) Oscillation (AO/NAO), the Pacific North American pattern (PNA), and the Pacific Decadal Oscillation (PDO) (McGuire *et al*., 2006). While the AO/NAO has turned to a more neutral state over the past decade following a period of enhanced westerly airflow associated with the positive phase of the AO/NAO, air temperatures have continued to show a warming trend. This trend coincides with the continued build-up of greenhouse gases, including atmospheric [CO_2], which are also thought to be an anthropogenic cause of increasing temperatures in this region (IPCC report). During the two time periods in this study, the later period, 1970–2000, is marked by a larger (44 ppm) increase in atmospheric [CO_2] and temperature, while the earlier period is marked by a smaller increases in atmospheric [CO_2] (10 ppm) and smaller increases in temperature (Fig. 1, Keeling *et al*., 1995). Our analyses indicated distinct differences in snow return and snow melt dates between the two time periods, with the earlier period marked by increased autumn temperatures and later snow return, whereas the later period was associated with warmer spring temperatures and earlier snow melt. Owing to the asymmetry in energy changes between pre- and post-snow-melt and pre- and post-snowfall, as discussed in 'Consideration of snow-cover variation in terms of both snow melt and snow return', it was precisely these difference in snow return and snow melt dates that resulted in a magnification of the atmospheric heating during the later period. Thus, our estimates of atmospheric heating due to changes in snow cover in the later period will likely to continue if air temperature trends in the spring continue to increase.

Changes in vegetation structure in relation to feedbacks in atmospheric heating

The large differences in atmospheric heating in vegetation types with large seasonal contrasts in albedo underscores the importance of considering those factors that may modify the cover type and surface albedo. Short-stature tundra vegetation is expected to be replaced, in part, by a greater density of taller shrubs and boreal forest in a warmer climate (Sturm *et al*., 2001). This expansion of the taller shrubs and trees could increase atmospheric heating through decreases in surface albedo (Bonan *et al*., 1992), but it might make these regions less sensitive to changes in snow-cover duration. We did not consider shifts in vegetation in this study as evidence suggests that, at present, areas with a lengthening snow-free season have contributed more strongly to atmospheric heating than have vegetation changes (Chapin *et al*., 2005). However, future shrub expansion in Alaska may contribute more to atmospheric heating than the changes in snow cover (Chapin *et al*., 2005), and would warrant careful consideration in studies that examine future trends in snow cover on atmospheric heating. The influence of the recent melting of sea-ice on albedo exerts additional positive feedback on atmospheric heating (Serreze & Francis, 2006).

Other changes in vegetation structure could act as a negative feedback (Chapin *et al*., 2000; McGuire *et al*., 2006). For example, increased fire activity in high-latitude forests associated with climate warming could exert a cooling effect, as vegetation with high winter albedo (bare ground and deciduous shrubs/trees) replaces evergreen forests with low winter albedo (Randerson *et al*., 2006). Likewise, increased timber harvest in high-latitude regions would increase albedo, as large stands of trees are replaced by bare ground and deciduous shrubs/trees (Bonan *et al*., 1992). However, albedo and biomass changes may not change in concert or linearly. That is, increases in the biomass of an aggrading forest may occur over a relatively long period of time while increases in albedo may occur rapidly after timber harvest or fire (Betts, 2000).

Consideration of forcings from atmospheric CO_2 concentrations

The positive feedbacks to atmospheric heating detected in this analysis are interesting to consider in light of negative feedbacks to atmospheric heating that may be induced by increases in carbon storage by the soils and vegetation of terrestrial regions. Currently, the increase in atmospheric heating due to increases in atmospheric CO_2 is approximately $3.7\,W\,m^{-2}$ (Myhre *et al.*, 1998). However, studies of forest inventory and satellite data collected between 1981 and 1999 have identified statistically significant ($P < 0.0001$) carbon gains in Eurasian boreal and North American temperate forests related to increased temperatures (Myneni *et al.*, 1997, 2001) that may offset heating due to increases in atmospheric CO_2 concentrations associated with fossil fuel emissions. Simulations with the TEM that included the effects of transient climate and increased atmospheric CO_2 concentrations indicate that between the years 1960 and 2000, the terrestrial regions between 60 and 90°N showed statistically significant gains in carbon uptake (Euskirchen *et al.*, 2006). During this time, the 60–90°N region switched from acting as a slight source of C during the 1960s (+0.05 Pg C, with a positive value indicating a C source) to a slight sink of CO_2 (−0.04 Pg C, with a negative value indicating a C sink). The offset of atmospheric heating by this small amount of increased C storage in recent decades is miniscule compared with the increases in atmospheric heating due to decreases in snow cover (see also Zhuang *et al.*, 2006).

Conclusion

Here, we considered the effects of changes in snow cover on energy feedbacks of northern high-latitude ecosystems in two warming periods of the 20th century. Our results suggest that during the recent 1970–2000 warming period, the heating of the atmosphere associated with changes in snow cover ($0.9\,W\,m^{-2}\,decade^{-1}$) was approximately three times that of the earlier 1910–1940 warming period ($0.3\,W\,m^{-2}$ decade). Furthermore, we found distinct differences in snow return and snow melt dates between the periods of analysis, with the earlier period marked by increased autumn temperatures and later snow return, whereas the later period was associated with warmer spring temperatures and earlier snow melt. Consequently, changes in atmospheric heating from 1970–2000 compared with 1910–1940 were magnified relative to changes in snow-cover duration because changes in the atmospheric heating of snow-covered and snow-free ecosystems were greater in spring than in autumn. Large differences in atmospheric heating were also noted between the vegetation types with high albedo contrast (e.g. tundra) and low albedo contrast (e.g. forests) even if they exhibited similar changes in snow-cover duration. It is important to continue to observe changes in snow cover and vegetation to determine how these and other changes of northern high-latitude ecosystems will influence the climate system.

Acknowledgements

Funds were provided by the National Science Foundation for the Arctic Biota/Vegetation portion of the 'Climate of the Arctic: Modeling and Processes' project (OPP-0327664). Joy Clein assisted with the model parameterization of the deciduous needleleaf forest. Catharine Copass Thompson and Jeff McAllister provided assistance with the vegetation map. John Walsh of the International Arctic Research Center offered valuable comments on an earlier draft of this manuscript.

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