Change in snow phenology and its potential feedback to temperature in the Northern Hemisphere over the last three decades

Shushi Peng1, Shilong Piao1,2,6, Philippe Ciais3, Pierre Friedlingstein4, Liming Zhou5 and Tao Wang3

1 Department of Ecology, College of Urban and Environmental Sciences, Peking University, Beijing 100871, People’s Republic of China
2 Institute of Tibetan Plateau Research, Chinese Academy of Sciences, Beijing 100085, People’s Republic of China
3 LSCE, UMR CEA-CNRS-UVSQ, CEA, L’Orme des Merisiers, F-91191 Gif-sur-Yvette, France
4 QUEST/Department of Earth Sciences, University of Bristol, Bristol BS8 1J, UK
5 Department of Atmospheric and Environmental Sciences, State University of New York, Albany, NY 12222, USA

E-mail: slpiao@pku.edu.cn

Received 25 October 2012
Accepted for publication 7 January 2013
Published 21 January 2013
Online at stacks.iop.org/ERL/8/014008

Abstract
Trends in the duration or extent of snow cover are expected to feedback to temperature trends. We analyzed trends in dates of onset and termination of snow cover in relation to temperature over the past 27 years (1980–2006) from over 636 meteorological stations in the Northern Hemisphere. Different trends in snow duration are observed over North America and Eurasia. Over North America, the termination date of snow cover remained stable during the 27 years, whereas over Eurasia it has advanced by 2.6 ± 5.6 d decade⁻¹. Earlier snow cover termination is systematically correlated on a year-to-year basis with a positive temperature anomaly during the snowmelt month with a sensitivity of −0.077 °C d⁻¹. These snow feedbacks to air temperature are more important in spring, because high net radiation is coupled with thin snow cover.

Keywords: climate change, climate feedback, global warming, snow phenology

Online supplementary data available from stacks.iop.org/ERL/8/014008/mmedia

1. Introduction

The widely distributed snow cover across the Northern Hemisphere (NH) during the winter season has been shown to be a sensitive indicator of climate change (IPCC 2007). As a consequence of current global warming, particularly over the high latitude regions (IPCC 2007), the spatial extent and the duration of snow cover in the NH have both decreased remarkably, and will continue to do so in the future, according to climate projections (IPCC 2007, Brown and Robinson 2011, Smith et al 2004). Both satellite measurements and in situ observations robustly document a persistent decreasing trend in snow cover extent (IPCC 2007, Groisman et al 1994, Serreze and Francis 2006, Dye 2002, Choi et al 2010) and a shortening of snow cover duration associated with the increase...
in temperature during winter and spring over the high latitude regions of NH (Smith et al. 2004, Choi et al. 2010, Kapnick and Hall 2010). For example, using multi-data sets from satellites and climate reanalysis products, Brown and Robinson (2011) reported that spring (March and April) snow cover extent over NH retreated by $-0.8$ million km$^2$ decade$^{-1}$ (7% in March and 11% in April) from 1970 to 2010. Similarly, using weekly snow cover dataset derived from NOAA satellites data (Robinson 1993), Choi et al. (2010) also found that full snow cover duration decreased by $5.3$ d decade$^{-1}$ from 1972 to 2007, mostly due to earlier snowmelt date. The decrease in snow cover and earlier snow melt in NH have important impacts on the hydrological cycle and biogeochemical cycles of terrestrial ecosystems (Smith et al. 2004, Stewart et al. 2005, Gripper et al. 2005). The spring stream-flow peak could shift earlier and affect the soil water storage in response to earlier snowmelt (Stewart et al. 2005). Moreover, changes in snow cover extent and snow melt timing have been shown to correlate with vegetation activity (Gripper et al. 2005, Peng et al. 2010) and to impact the carbon balance of terrestrial ecosystems (Monson et al. 2006, McDonald et al. 2004, Zhou et al. 2001, Jonas et al. 2008). Therefore, it is essential to understand the processes that control snow cover changes (IPCC 2007, Brown and Robinson 2011).

Previous studies suggested that the decrease in snow cover extent and duration is a direct response to the increasing temperature (Brown and Robinson 2011, Stone et al. 2002, Chapin et al. 2005). During the period 1922–2010, air temperature could explain $\sim$50% of the interannual variance in NH spring snow cover extent (Brown and Robinson 2011), and the sensitivity of spring snow cover extent to temperature is estimated as $-0.8$ to $-1.0$ million km$^2$ °C$^{-1}$ (Brown and Robinson 2011). Hence, changes in snow cover extent and duration are indicators of warming (IPCC 2007, Chapin et al. 2005). But in turn, changes in snow cover extent and duration also feedback on temperature. Changes in snow cover and snow thickness alter the surface albedo (Chapin et al. 2005, Euskirchen et al. 2007), the surface energy partitioning (Groisman et al. 1994), and the hydrological cycle (Stewart et al. 2005, Hamlet et al. 2005). The decrease in surface albedo due to decrease in snow cover fractional area and duration could induce more absorbed solar radiation at the surface, which in turn amplify the increase in temperature. These processes are estimated to act as a robust positive feedback, i.e. to further amplify the warming trends in pan-Arctic regions (Kapnick and Hall 2010, Déry and Brown 2007, Chapin et al. 2005, Euskirchen et al. 2007). This feedback is usually referred as snow albedo feedback (Bony et al. 2006, Hall and Qu 2006, Fernandez et al. 2009).

Stone et al. (2002) estimated that an 8 d advance of snow melt date led to additional $\sim64$ MJ m$^{-2}$ of absorbed solar radiation ($\sim2$ W m$^{-2}$ feedback forcing on an annual basis) at the surface of Barrow station from 1966 to 2000. Using satellite-based snow cover and planetary albedo data products, Groisman et al. (1994) estimated an increase in net radiation of 0.9 W m$^{-2}$ due to the albedo decrease from decreasing snow cover extent over northern extratropical land regions from 1979 to 1990. Comparing the observed snow albedo feedback with the simulation results of 17 IPCC AR4 climate models, the sensitivity of surface albedo to temperature is the main uncertainty in the models (0.4–1.4% K$^{-1}$) (Hall and Qu 2006). Thus, quantifying and understanding the sensitivity of surface albedo to temperature in the snow albedo feedback is critical for assessing and projecting the future climate.

In this study, using daily observations of snow cover and monthly mean air temperature from 636 meteorological stations over NH, we first characterize the spatial patterns of change in snow cover duration and the snow onset and end dates in different regions of the NH, and then estimate the sensitivity of snow disappearance date to surface air temperature and the feedback, defined in this study as the sensitivity of surface air temperature to snow disappearance date during the period 1980–2006. Moreover, we separate the contribution of snow albedo feedback to the observed increase in surface air temperature during the snow melt season in spring.

2. Datasets and methods

2.1. Datasets

Table 1 summarizes the datasets used in this study. Daily snow depth and surface air temperature data from August 1979 to July 2006 in China, Europe, Russia, Canada and USA were obtained from the National Meteorological Information Center of the China Meteorological Administration (Peng et al. 2010), European Climate Assessment & Dataset (ECA&D) (available at http://ecad.knmi.nl/dailydata/predefinedseries.php), Russian Research Institute of Hydrometeorological Information—World Data Center (RIHMI-WDC) (Bulygina et al. 2011), the National Climate Data and Information Archive of Environment Canada (available at http://climate.weatheroffice.gc.ca/prods_servs/index_e.html) and the United States Historical Climatology Network (USHCN) (available at http://cdiac.ornl.gov/epubs/ndp/ushcn/ushcn.html), respectively. The albedo dataset used in this study was from EOS-Terra-MODIS MOD43C3 Nadir BRDF-Adjusted Reflectance (NBAR) 2001–2006 products, at a spatial resolution of 0.05°×0.05° and 8 d interval including black sky albedo (BSA) and white sky albedo (WSA) over short-wave broadband (0.3–5.0 µm) (Schaf et al. 2002). WSA shows similar results to BSA, so we used BSA to illustrate the results here. The bias in MODIS albedo data (Collection V005) is mostly less than 5% (Liu et al. 2009). The incoming short-wave radiation dataset used in this study was from ERA-Interim reanalysis produced by European Centre for Medium-Range Weather Forecasts (ECMWF) over the period 2001–2006, at $\sim0.7^\circ\times\sim0.7^\circ$ spatial resolution and 6 h temporal resolution, with the bias of $\sim2$ W m$^{-2}$ (Dee et al. 2011).

2.2. Analyses

To avoid the impact of ephemeral snow on the ground on the snow cover duration, the onset date of snow cover for year (t) at each meteorological station is defined as the first continuous five days on which snow was observed to cover the ground.
surface in the fall of the year \((t - 1)\); the end date of snow cover is defined as the last continuous five days when snow cover was observed in the spring of the year \(t\). The snow cover duration is defined as the number of days from the onset date of snow cover to the end date of snow cover for the year \(t\). For any site, the year \(t\) was omitted from the analysis if there were one or more days without snow depth data during the period from 1 August of year \((t - 1)\) to 31 July of year \((t)\). Furthermore, in order to reduce uncertainties caused by the short time period of observation, we consider only stations for which at least 20 years of daily snow depth data are available during the study period. Among 636 meteorological stations, which have at least 20 years of daily snow depth data, only 575 stations have at least 20 years of both daily snow depth and monthly air temperature data during the study period 1980–2006. Maximum snow depth during snow season is calculated as the maximum snow depth record from the onset date of snow cover to the end date of snow cover. We referred spring and fall as the snowmelt (February–July) and onset (September–January) season.

To map the spatial distribution of the trend of the end date of snow cover \(D_E\), we applied the trend detection methodology from Weatherhead et al (1998) which consider the autocorrelation of the underlying noise processes (assuming noise is autoregressive of the order of 1), at each of the 636 meteorological stations with at least 20 years of daily snow depth data during the period 1980–2006. We similarly calculated the trend of the onset date of snow cover \(D_O\), the duration of snow cover and the temperature at each site.

For the sensitivity of \(D_E\) to temperature, to remove the effects of snow depth on \(D_E\), we regressed the end date \(D_E\) as dependent variable with average air temperature during the month before snowmelt \((T_{M-1})\) and maximum snow depth \((SD_M)\) as independent variables.

\[
D_E = a \times T_{M-1} + b \times SD_M + c \tag{1}
\]

where \(a\) and \(b\) are regression coefficients, and \(c\) is the residual error. \(D_E\) mainly depends on the temperature before snowmelt and the maximum snow depth to melt (see section 3). We defined the regression coefficient of \(T_{M-1}\) \((a)\) as the sensitivity of \(D_E\) to temperature.

To examine the effect of positive feedback of earlier snowmelt on temperature, we performed a multiple linear regression (equation (2)) using average air temperature during snowmelt month \((T_M)\) as the dependent variable, and \(D_E\) and average air temperature during the month before snowmelt \((T_{M-1})\) as independent variables.

\[
T_M = s \times D_E + p \times T_{M-1} + q \tag{2}
\]

where \(s\) and \(p\) are regression coefficients, and \(q\) is the residual error. We assumed that interannual variations of \(T_M\) come from general atmosphere circulation and the local snow albedo feedback, which are reflected by the term \(p \times T_{M-1}\) and \(s \times D_E\), respectively. The slope of the regression, \(s\), defines the sensitivity of temperature to \(D_E\).

Similarly, for the sensitivity of temperature to \(D_O\), we performed the multiple linear regression as equation (2), but using the average air temperature during snow onset month.

---

**Table 1.** Datasets used in this study.

| Datasets                      | Method                                      | Spatial domain | Time span   | Temporal resolution | Number of stations/spatial resolution | Source/reference                                                                 |
|-------------------------------|---------------------------------------------|----------------|-------------|--------------------|---------------------------------------|--------------------------------------------------------------------------------|
| Snow depth and surface air temperature | Ground-based meteorological station observation | China          | 1979–2006   | Daily              | 108                                   | National Meteorological Information Center of the China Meteorological Administration (Peng et al 2010) |
|                               |                                             | Europe         | 1979–2006   | Daily              | 108                                   | European Climate Assessment & Dataset (ECA&D) (available at http://ecad.knmi.nl/dailydata/predefinedseries.php) |
|                               |                                             | Russia         | 1979–2006   | Daily              | 89                                    | Russian Research Institute of Hydrometeorological Information World Data Center (RIHMI-WDC) (Bulygina et al 2011) |
|                               |                                             | Canada         | 1979–2006   | Daily              | 130                                   | National Climate Data and Information Archive of Environment Canada (available at http://climate.weatheroffice.gc.ca/ prods_servs/index_e.html) |
|                               |                                             | United States  | 1979–2006   | Daily              | 201                                   | United States Historical Climatology Network (USHCN) (available at http://cdiac.ornl.gov/epubs/ndp/ushcn/ushcn.html) |
| Albedo                        | Satellite                                   | Global         | 2001–2006   | 8 d                | 0.05° × 0.05°                         | www.modis.bu.edu/birds/product.html (Schaaf et al 2002) |
| Incoming short-wave radiation | Reanalysis                                   | Global         | 2001–2006   | 6 h                | ∼0.7° × ∼0.7°                        | European Centre for Medium-Range Weather Forecasts (ECMWF) (Dee et al 2011) |
as the dependent variable, and $D_O$ and average air temperature during the month before snow onset ($T_{O-1}$) as independent variables as equation (3)

$$T_O = s \times D_O + p \times T_{O-1} + q,$$

where $s$ and $p$ are regression coefficients, and $q$ is the residual error. The slope of the regression, $s$, defines the sensitivity of temperature to $D_O$.

For the incoming short-wave (SW) radiation at surface at $D_E$ or $D_O$ in each station, we used average daily incoming SW radiation at $D_E$ or $D_O$ during the period 2001–2006 in the pixel (6 h, $\sim0.7^\circ \times \sim0.7^\circ$) where the station located. For the surface albedo, we resample the station surface albedo from MODIS MOD43C3 (8 d, 0.05° × 0.05°). For each station, we used maximum and minimum black sky albedo (BSA) and white sky albedo (WSA) from MODIS MOD43C3 during the month of snowmelt as the surface albedo before and after snowmelt (figure S1 available at stacks.iop.org/ERL/8/014008/mmedia), respectively. Then we took the difference in BSA before and after snowmelt as $\delta$BSA. The product of incoming SW radiation during snowmelt (snow onset) by $\delta$BSA is the difference in absorbed short-wave energy (ASR) between snow free and snow covered surface ($\delta$ASR).

3. Results and discussion

3.1. Changes in the onset (end) date and duration of snow cover from 1980 to 2006

Figures 1(A) and (B) show the spatial patterns of mean onset and end date of snow cover during the period 1980–2006. There are clear latitudinal gradient patterns for the onset (earlier with higher latitude) and end (later with higher latitude) date of snow cover (figures 1(A) and (B)). Then the mean snow cover duration shows longer snow duration period with higher latitude (figure 1(C)). We first applied the trend detection method (Weatherhead et al 1998) to the time series of the onset and end date of the snow cover in each year, and of the snow cover duration at each station (figure 2). A decrease in snow cover duration is found for 63% of the stations (figure 2(C)). This result agrees with the observation from Scanning Multi-channel Microwave Radiometer (SMMR) satellite data with global coverage at 25 km resolution and 2 d revisit (Choi et al 2010, Takala et al 2009). The rate of shrinking of snow duration is greater than five days per decade at more than 34% of the 636 stations (figure 2(C)). Interestingly, North America and Eurasia show different trends for the end date of snow cover in spring. A trend towards an earlier end date of snow cover is found at 70% of the Eurasian sites, with 28% of the sites showing an advance of snowmelt in spring as fast as five days per decade. By contrast, the trends in the end date of snow cover over North America exhibit a fragmented spatial pattern. At the continental scale, the end date of snow cover over the last 27 years shows a rapid advance for Eurasia ($-2.6 \pm 5.6$ d decade$^{-1}$) and virtually no trend for North America ($0.1 \pm 5.8$ d decade$^{-1}$). The onset date of the first snow shows a trend towards later dates in Eurasia ($1.3 \pm 4.9$ d decade$^{-1}$) and North America ($1.1 \pm 4.9$ d decade$^{-1}$). These results are consistent with the trends of vegetation phenology estimated by satellite observation and some terrestrial biogeochemical cycle models (Smith et al 2004, Tucker et al 2001, Piao et al 2008, Jeong et al 2011). The green-up of vegetation in spring for many ecosystems north of 25°N was modeled to follow a trend.
Figure 2. Trends in (A) the onset date of snow cover (d decade$^{-1}$), (B) the end date of snow cover (d decade$^{-1}$), and (C) snow cover duration (d decade$^{-1}$) from 1980 to 2006 over 636 meteorological stations in the Northern Hemisphere. The insets show the frequency distributions of corresponding trends. Black-edged circles indicates that the trend is statistically significant ($P < 0.05$).

towards earlier dates in Eurasia (Piao et al 2008, Jeong et al 2011), in agreement with satellite observation of vegetation reflectance (Zhou et al 2001).

3.2. Sensitivity of the end date of snow cover to temperature

We hypothesize that the geographically different trends in snow cover phenology between North America and Eurasia for the end date of snow cover, are associated with local temperature trends. From co-located temperature records at 575 snow measurement stations (see datasets and methods), we found that in the seasons of snow melt (February–July), 85% of the Eurasian meteorological stations indeed show an increase in temperature during the month of snowmelt ($T_M$) and $T_M$ increased by $0.21 \pm 0.18 \, ^\circ\text{C}\,\text{decade}^{-1}$ for all Eurasian stations, whereas only 49% of the North American stations show an increase in $T_M$ and at a smaller rate of $0.00 \pm 0.21 \, ^\circ\text{C}\,\text{decade}^{-1}$ for all North American stations. During the fall (seasons of snow accumulation beginning), however, air temperature appears to increase at a similar rate over North America ($0.10 \pm 0.15 \, ^\circ\text{C}\,\text{decade}^{-1}$) as over Eurasia ($0.13 \pm 0.14 \, ^\circ\text{C}\,\text{decade}^{-1}$), in agreement with the hypothesis of a co-variation between temperature and the presence of snow.

In addition to spring air temperature, we also hypothesize that the end date of snowmelt is also related to snow depth, a proxy for the amount of snow to be melted (Stone et al 2002). Among the 636 stations in this study, 95% of them show a positive correlation of the end date of snowmelt with the maximum snow depth during the snow season (statistically significant at 49% of the 636 stations). This corroborates the work by Stone et al (2002) who reported that snow water equivalent influences the snowmelt date at Barrow station. We found that the spatial distribution of the trend in maximum snow depth during the snow season can only explain a limited (7.0%), but statistically significant ($P < 0.001$) portion of the observed distribution in the trend in end date of snowmelt over the 636 stations.

To estimate the coupling between spring snow disappearance and air temperature, we first performed a correlation analysis between the end dates of snow cover each year and $T_{M-1}$ (monthly air temperature during the month before snowmelt) across 575 stations where at least 20 years of daily snow cover and collocated air temperature data were available during the study period. A negative correlation between the end date of snow cover and $T_{M-1}$ was obtained at more than 94% of the 575 stations (statistically significant at 40% of the stations), suggesting that spring air temperature plays a critical role in controlling the timing of snowmelt. Because both maximum snow depth during snow season and spring air temperature can influence the end date of snow cover, equation (1) was applied to estimate the sensitivity of the end date of snow cover to $T_{M-1}$ after controlling the effect of snow depth on the end date of snow cover for each station (partial correlation analysis). We found that, for a given increase of $T_{M-1}$ of 1°C, the end date of snow cover advances on average by $1.6 \pm 1.8 \, \text{d}$ (figure 3(A)). In addition to the impact of warmer spring temperatures on the end date of snow cover, a feedback of snowmelt on air temperature is also expected (Groisman et al 1994, Déry and Brown 2007, Hall and Qu 2006). The magnitude of this feedback is analyzed below.
3.3. Sensitivity of spring (autumn) temperature to the end (onset) date of snow cover

In this section, we made an attempt to quantify the feedback of snowmelt on spring temperature. To do so, we established a relationship between the end date of snow cover ($D_E$) and temperature during the month of snowmelt ($T_M$) for each station (retaining only sites with at least 20 years of daily snow cover and air temperature data from 1980 to 2006). We found that 98% of the stations exhibited a negative correlation between $T_M$ and $D_E$ (statistically significant at 67% of stations). At 69% of the stations, $T_M$ shows an even stronger negative correlation with $D_E$ than with $T_{M-1}$. We performed multiple linear regression analysis using $T_M$ as the dependent variable and both $D_E$ and $T_{M-1}$ as independent variables (predictors) to further remove the effect of autocorrelation of temperature between two consecutive spring months (see datasets and methods). The result was that changes in $D_E$ are significantly and negatively correlated with $T_M$ at 332 of the 575 NH stations, when controlling for the effect of $T_{M-1}$. The probability that this occurs purely by chance is estimated to be less than 0.1% if all station records are assumed to be independent. The linear regression slope, or sensitivity, of spring temperature to $D_E$ is $-0.077 \pm 0.067^\circ C \ d^{-1}$ (figure 3(B)). This sensitivity can be interpreted as a measure of the snow-temperature feedback. In other words, an earlier disappearance of snow cover by 1 d could result in an additional warming of monthly temperature of $0.077 \pm 0.067^\circ C$.

Likewise, in the fall (onset of snow) we found positive correlations between the onset date of snow cover ($D_O$) and the temperature during the month of snow onset ($T_O$) in 98% of the stations (statistically significant at 54% of the stations). In other words, a colder autumn coincides with an earlier snow onset. When we remove the effect of temperature during the month preceding snow onset ($T_{O-1}$), the sensitivity of $T_O$ to $D_O$ is $0.068 \pm 0.051^\circ C \ d^{-1}$ (figure 3(C)), a little smaller in absolute value than the sensitivity of $T_M$ to $D_E$ ($-0.077 \pm 0.067^\circ C \ d^{-1}$). This result suggests a small feedback of snow cover on local temperature in autumn, a period of the year whose availability is controlled by albedo) is smaller than that in snowmelt season (spring).

In addition to temporal relationships at each site, we also found a negative (positive) correlation between trends in the end (onset) date of snow cover and trends in spring (autumn) temperature across the sites. We regressed the trends in $T_M$ ($T_O$) against the trends in $D_E$ ($D_O$) across the 575 stations, respectively (figure 4). The key result is that the trends in $T_M$ ($T_O$) are negatively (positively) associated with the trends in $D_E$ ($D_O$) across the sites with $R = -0.27, P < 0.001 (R = 0.22, P < 0.001)$. After removing the effect of the trend in $T_{M-1}$ ($T_{O-1}$), there is still a significant residual negative partial correlation between the trend in $T_M$ ($T_O$) and the trend in $D_E$ ($D_O$) across the stations ($R = 0.20, P < 0.001; R = -0.20, P < 0.001$; figure 4). The spatial sensitivity of temperature to $D_E$ ($D_O$), given by the linear slope between the $T_M$ ($T_O$) and $D_E$ ($D_O$) trend is $-0.007^\circ C \ d^{-1}$ (0.008$^\circ C \ d^{-1}$), which is smaller than the temporal sensitivity established above ($-0.077\pm0.067^\circ C \ d^{-1}$ for the sensitivity of $T_M$ to $D_E$ and $0.068 \pm 0.051^\circ C \ d^{-1}$ for the sensitivity of $T_O$ to $D_O$, respectively).

The sensitivity of $T_M$ ($T_O$) to $D_E$ ($D_O$) depends among others, on the amount of extra absorbed incoming short-wave
Figure 4. Relationship between (A) trend in the onset date of snow cover (DO) (d yr$^{-1}$) and trend in temperature during the month of snow onset (TO) ($^{\circ}$C yr$^{-1}$) and between (B) trend in the end date of snow cover (DE) (d yr$^{-1}$) and trend in temperature during the month of snowmelt (TM) ($^{\circ}$C yr$^{-1}$) across 575 meteorological stations in the Northern Hemisphere. Red and green filled circles indicate meteorological stations from Eurasia (EA) and North America (NA), respectively.

Figure 5. Relationship between the sensitivity of temperature during the month of snowmelt (TM) to the end date of snow cover (DE) and (A) incoming short-wave (SW) radiation at DE, (B) the difference in albedo (black sky albedo (BSA)) before and after snowmelt ($\delta$BSA), and (C) the difference in absorbed short-wave radiation (ASR) after and before snowmelt at surface ($\delta$ASR) across the 575 stations with matching albedo values from MODIS.

(SW) radiation due to earlier snowmelt (later snow onset), which is the product of incoming SW radiation during snowmelt (snow onset) and the difference of albedo between snow free and snow covered surfaces. We quantified two important driving factors that are hypothesized to control the sensitivity of TM (TO) to DE (DO): short-wave incoming radiation, and surface albedo. Firstly, we correlated the sensitivity of TM to DE to daily short-wave incoming radiation across the 575 selected stations. Figure 5(A) shows that the sensitivity of TM to DE is significantly and negatively correlated with incoming SW radiation at the end date of snow cover ($R = -0.29, P < 0.001$). Secondly, albedo is the other factor that can feedback on temperature during snowmelt as discussed in previous studies (Groisman et al 1994, Bony et al 2006, Hall and Qu 2006, Fernandes et al 2009, Qu and Hall 2006, Flanner et al 2011). An earlier disappearance of snow influences temperature locally by diminishing surface albedo, causing more incoming SW radiation to be absorbed by the surface (Hall and Qu 2006). Here, we used albedo (black sky albedo (BSA)) from EOS-Terra-MODIS (MODerate-resolution Imaging Spectrometer) MCD43C3 version 5 products over the period 2001–2006 separated between snow free and snow cover ($\delta$BSA) to estimate the snow albedo change around the snowmelt (see section 2 and figure S1 available at stacks.iop.org/ERL/8/014008/mmedia). Figure 5(B) shows the relationship between the temporal sensitivity of TM to DE (see above) and $\delta$BSA across the 575 stations for which there is matching albedo data over the period 2001–2006 ($R = -0.44, P < 0.001$). We found that the temporal sensitivity of TM to DE is significantly and negatively correlated with both incoming SW radiation and $\delta$BSA, verifying the larger difference in absorbed energy before and after snowmelt associated with larger snow albedo feedback on temperature (figures 5(A) and (B)). We took the product of incoming SW radiation by $\delta$BSA to present
the extra absorbed energy by surface because of change in $D_E (\delta ASR)$, which exhibited a stronger relationship with the \textit{temporal sensitivity} of $T_M$ to $D_E (R = -0.50, P < 0.001)$ than that using incoming SW radiation or $\delta BSA$ alone ($R = -0.29, P < 0.001; R = -0.44, P < 0.001$).

In addition, the \textit{temporal sensitivity} of $T_O$ to $D_O$ is much less correlated with incoming SW radiation during snow onset and $\delta ASR$ ($R = -0.21, P < 0.001; R = 0.06, P = 0.121$; figure S2 available at stacks.iop.org/ERL/8/014008/mmmedia). This may be related to the smaller incoming SW radiation during snow onset season (incoming SW radiation in 97% of stations smaller than 10 MJ d$^{-1}$) and $\delta ASR$ is limited ($\delta ASR$ in 97% of stations smaller than 2 MJ d$^{-1}$). However, the incoming SW radiation during snowmelt season (spring) exceeds 10 MJ d$^{-1}$ over most of the stations and $\delta ASR$ exceeds 2 MJ d$^{-1}$ over 51% of the stations. The albedo also has influence on the sensitivity of $T_O$ to $D_O$, but the explanation is limited ($R = 0.30, P < 0.001$; figure S2 available at stacks.iop.org/ERL/8/014008/mmmedia).

The average \textit{temporal sensitivity} in temperate regions (south of 50°N) ($-0.059 \pm 0.055 \ ^\circ \text{C} \ d^{-1}, n = 358$) is larger than that in boreal regions (north of 50°N) ($-0.106 \pm 0.075 \ ^\circ \text{C} \ d^{-1}, n = 217$) ($P < 0.001$). This may be related to the fact that a higher incoming SW radiation could be available during snowmelt in boreal regions ($16.1 \pm 4.1 \ \text{MJ} \ d^{-1}$) than in temperate regions ($14.5 \pm 3.8 \ \text{MJ} \ d^{-1}$) over NH ($P < 0.001$), considering the later snowmelt date in higher latitude (figure S3 available at stacks.iop.org/ERL/8/014008/mmmedia) (Choi et al 2010). In addition, $\delta BSA$ in boreal regions (0.17 ± 0.10) is also higher than that in temperate regions (0.14 ± 0.1). This could explain why the magnitude of snow albedo feedback is larger in the high latitude region over NH (Serreze and Francis 2006, Chapin et al 2005, Euskirchen et al 2007, Bony et al 2006).

At Barrow station (latitude is about 70°N), Stone et al (2002) found that temperature in June increased by 0.9 °C when snowmelt date advanced by 14 d (table 2 in their paper). In other words, the sensitivity of temperature to $D_E$ is $-0.064 \ ^\circ \text{C} \ d^{-1}$ at this site, which is a little smaller than the average result ($-0.092 \pm 0.054 \ ^\circ \text{C} \ d^{-1}$) at 12 stations located between 69°N and 71°N in this study. Based on the seasonal cycles of satellite-derived albedo and air temperature, the sensitivity of surface albedo to temperature was estimated as $-0.93 \pm 0.06\% \text{ K}^{-1}$ and $-1.06 \pm 0.08\% \text{ K}^{-1}$ in Fernandes et al (2009) and Hall and Qu (2006) over the regions north of 30°N, respectively. In this study, the sensitivity of surface albedo to air temperature is estimated as the product of the sensitivity of $D_E$ to temperature by $\delta BSA$ of $-0.90 \pm 0.86\% \text{ K}^{-1}$. In terms of radiative forcing, we used the product of sensitivity of $D_E (D_O)$ to temperature by $\delta ASR$ as the magnitude of snow albedo feedback, which are $0.13 \pm 0.13 \ \text{W m}^{-2} \text{ K}^{-1}$ and $0.06 \pm 0.09 \ \text{W m}^{-2} \text{ K}^{-1}$ in spring and autumn, respectively. The total annual snow albedo feedback (spring plus autumn) over the land of Northern Hemisphere found in this study is $0.19 \pm 0.17 \ \text{W m}^{-2} \text{ K}^{-1}$, while the total cryosphere albedo feedbacks including snow and ice over NH was estimated to lie in the range of $0.3-1.1 \ \text{W m}^{-2} \text{ K}^{-1}$ from satellite observation, and to average at $0.25 \pm 0.17 \ \text{W m}^{-2} \text{ K}^{-1}$ from 18 CMIP3 climate models (IPCC 2007, Flanner et al 2011). The estimate of our study, based on \textit{in situ} records is comparable to these earlier estimates, although our results only include the local albedo feedback and are limited to the amount of \textit{in situ} stations and their spatial distribution. The uncertainty in our estimate is larger, however, because it includes the variation across the 575 stations.

In this study, we used the extra absorbed energy at surface due to earlier snow disappearance (later snow onset) to explain the magnitude of the sensitivity of temperature to the end (onset) date of snow cover, assuming that the sensible heat flux allocation is related to absorbed energy at surface. Actually, earlier snowmelt can impact air temperature through changing the hydrological cycle. Snowmelt increases surface runoff in the beginning of the snowmelt period, but earlier snowmelt potentially leads to decreased soil water availability by the end of the snowmelt period (Hamlet et al 2005), particularly if frozen soils prevent snowmelt from recharging the soil moisture reservoir (Dunne runoff) (Dunne and Black 1970). As a consequence, an increase in the exchange of sensible heat between the earth surface and the atmosphere is expected; this will further contribute to an increase in $T_M$ (Seneviratne et al 2006, Tchebakova et al 2002, Chapin et al 2008). In addition, we did not consider the cloud variation which could impact the solar radiation at the surface, and then impact the snow albedo feedback (Bony et al 2006, Qu and Hall 2006, Flanner et al 2011). More complex drivers such as sensible and latent heat flux allocation and cloud impacts on the snow albedo feedback should be included in future modeling studies.

4. Conclusions

From 1980 to 2006, at the continental scale, the end date of snow cover over the last 27 years shows a rapid advance for Eurasia ($-2.6 \pm 5.6 \ \text{d decade}^{-1}$) and virtually no trend for North America ($0.1 \pm 5.8 \ \text{d decade}^{-1}$). The onset date of the first snow cover shows a trend towards later dates in both Eurasia ($1.3 \pm 4.9 \ \text{d decade}^{-1}$) and North America ($1.1 \pm 4.9 \ \text{d decade}^{-1}$). The differences in snow cover trends between North America and Eurasia are associated with local temperature trends. By the end of this century, a global increase in surface air temperature will result from increasing concentrations of greenhouse gases in the atmosphere (IPCC 2007). Our analyses of available records from meteorological stations indicate that snow cover in the NH is very sensitive to rising temperature, and that there could be a positive feedback of snow cover change to spring temperature. When we apply this empirical sensitivity of temperature to snow disappearance date ($-0.077 \ ^\circ \text{C} \ d^{-1}$) to the average observed trend in snow disappearance date ($-1.3 \ \text{d decade}^{-1}$) across the 575 stations where at least 20 years of daily snow cover and air temperature data are available from 1980 to 2006, we predict a contribution of snow cover feedback to spring warming to be $0.10\% \text{ C} \ \text{d}^{-1}$. This rate accounts for 32% of the observed linear trend in spring temperature from 1980 to 2006 (0.31 \ ^\circ \text{C} \ \text{d}^{-1})$. Furthermore, applying the future northern land surface (north of 30°N) spring warming of
1.0–5.3 °C predicted by IPCC climate models to the average temperature sensitivity of the end date of snow cover observed in this study (1.6 °C–1) gives a projected advance of 1.6–8.5 d by the end of this century, which will probably further amplify spring warming by 0.12–0.65 °C. Note that this does not account for further atmospheric circulation feedbacks. Our observational estimate of the snow-temperature feedback may contain very useful information that can be used to evaluate coupled land-surface–atmosphere models, and to improve the quantification of snow–climate feedbacks.

Acknowledgments

We thank Dr Olga N Bulygina for the Russian meteorological station datasets and Dr J H Gash for English language editing. This study was supported by the National Natural Science Foundation of China (grant 41125004 and 30970511), and CARBONES EU F7 foundation (242316).

References

Bony S et al 2006 How well do we understand and evaluate climate change feedback processes J. Clim. 19 3445–82
Brown R D and Robinson D A 2011 Northern Hemisphere spring snow cover variability and change over 1922–2010 including an assessment of uncertainty Cryosphere 5 219–29
Bulygina O N, Groisman P Y, Razuvaev V N and Korshunova N N 2011 Changes in snow cover characteristics over Northern Eurasia since 1966 Environ. Res. Lett. 6 045204
Chapin F S et al 2005 Role of land-surface changes in Arctic summer climate Science 310 657–60
Chapin F S, Randerson J T, McGuire A D, Foley J A and Field C B 2008 Changing feedbacks in the climate–biosphere system Front. Ecol. Environ. 6 313–20
Choi G, Robinson D A and Kang S 2010 Changing Northern Hemisphere snow seasons J. Clim. 23 5305–10
Dee D P et al 2011 The ERA-Interim reanalysis: configuration and performance of the data assimilation system Q. J. R. Meteorol. Soc. 137 553–97
Déry S J and Brown R D 2007 Recent Northern Hemisphere snow cover extent trends and implications for the snow–albedo feedback Geophys. Res. Lett. 34 L22504
Dunne T and Black R D 1970 Partial area contributions to storm runoff in a small New England watershed Water Resour. Res. 6 1296–311
Dye D G 2002 Variability and trends in the annual snow-cover cycle in Northern Hemisphere land areas, 1972–2000 Hydrol. Process. 16 3065–77
Euskirchen E S, McGuire A D and Chapin F P III 2007 Energy feedbacks of northern high-latitude ecosystems to the climate system due to reduced snow cover during 20th century warming Glob. Change Biol. 13 2425–38
Fernandez R, Zhao H, Wang X, Key J, Qu X and Hall A 2009 Controls on Northern Hemisphere snow albedo feedback quantified using satellite Earth observations Geophys. Res. Lett. 36 L21702
Flanner M G, Shell K M, Barlage M, Perovich D K and Tschudi M A 2011 Radiative forcing and albedo feedback from the Northern Hemisphere cryosphere between 1979 and 2008 Nature Geosci. 4 151–5
Grippa M, Kergoat L, Le Toan T, Mognard N M, Delbart N, L. Hermitte J and Vicente-Serrano S M 2005 The impact of snow depth and snowmelt on the vegetation variability over central Siberia Geophys. Res. Lett. 32 L21412
Groisman P Y, Karl T R and Knight R W 1994 Observed impact of snow cover on the heat balance and the rise of continental spring temperatures Science 263 198–200
Hall A and Xu Q 2006 Using the current seasonal cycle to constrain snow albedo feedback in future climate change Geophys. Res. Lett. 33 L03502
Hamlet A F, Mote P W, Clark M P and Lettenmaier D P 2005 Effects of temperature and precipitation variability on snowpack trends in the western United States J. Clim. 18 4545–61
IPCC 2007 Climate Change 2007: The Physical Science Basis (Cambridge: Cambridge University Press)
Jonas T, Rixen C, Sturm M and Stoeckli V 2008 How alpine plant growth is linked to snow cover and climate variability J. Geophys. Res. 113 G03011
Jeong S-J, Ho C-H, Gim H-J and Brown M E 2011 Phenology shifts at start vs. end of growing season in temperate vegetation over the Northern Hemisphere for the period 1982–2008 Glob. Change Biol. 17 2385–99
Kapnick S and Hall A 2010 Observed climate–snowpack relationships in California and their implications for the future J. Clim. 23 3446–56
Liu J, Schaf C, Strahler A, Jiao Z, Shuai Y, Zhang Q, Roman M, Augustine J A and Dutton E G 2009 Validation of moderate resolution imaging spectroradiometer (MODIS) albedo retrieval algorithm: dependence of albedo on solar zenith angle J. Geophys. Res. 114 D01106
McDonald K C, Kimball J S, Njoku E, Zimmermann R and Zhao M S 2004 Variability in springtime thaw in the terrestrial high latitudes: monitoring a major control on the biospheric assimilation of atmospheric CO2 with spaceborne microwave remote sensing Earth Interact. 8 I–23
Monson R K, Lipson D L, Burns S P, Turnipseed A A, Delany A C, Williams M W and Schmidt S K 2006 Winter forest soil respiration controlled by climate and microbial community composition Nature 439 711–4
Peng S S, Piao S L, Ciais P, Fang J Y and Wang X H 2010 Change in winter snow depth and its impacts on vegetation in China Glob. Change Biol. 16 3004–13
Piao S L et al 2008 Net carbon dioxide losses of northern ecosystems in response to autumn warming Nature 451 49–52
Qu X and Hall A 2006 Assessing snow albedo feedback in simulated climate change J. Clim. 19 2617–30
Robinson D A 1993 Hemispheric snow cover from satellites Annals Glaciol. 17 367–71
Schaaf C B et al 2002 First operational BRDF, albedo nadir reflectance products from MODIS Remote Sens. Environ. 83 135–48
Seneviratne S I, Luthi D, Litschi M and Schar C 2006 Land–atmosphere coupling and climate change in Europe Nature 433 205–9
Serreze M C and Francis J A 2006 The arctic amplification debate Clim. Change 76 241–64
Smith N V, Saatchi S S and Randerson J T 2004 Trends in high latitudes: monitoring a major control on the biospheric assimilation of atmospheric CO2 with spaceborne microwave remote sensing Earth Interact. 8 I–23
Stewart I T, Cayan D R and Dettinger M D 2005 Changes toward earlier spring warming J. Clim. 18 3241–51
Stone R S, Dutton E G, Harris J M and Longenecker D 2002 Earlier spring snowmelt in northern Alaska as an indicator of climate change J. Geophys. Res. 107 D08089
Takala M, Pulliainen J, Miettikangas S J and Koskinen J T 2009 Detection of snowmelt using spaceborne microwave radiometer data in Eurasia from 1979 to 2007 IEEE Trans. Geosci. Remote Sens. 47 2996–3007
Tchekarev N M, Kolle O, Zolotoukhine D, Arneath A, Styles J M, Vygodskaya N N, Schulze E D, Shibistova O and Lloyd J 2002
Inter-annual and seasonal variations of energy and water vapour fluxes above a *Pinus sylvestris* forest in the Siberian middle taiga *Tellus* 54 537–51

Tucker C J, Slayback D A, Pinzon J E, Los S O, Myneni R B and Taylor M G 2001 Higher northern latitude normalized difference vegetation index and growing season trends from 1982 to 1999 *Int. J. Biometeorol.* 45 184–90

Weatherhead E C et al 1998 Factors affecting the detection of trends: statistical considerations and applications to environmental data *J. Geophys. Res.* 103 17149–61

Zhou L M, Tucker C J, Kaufmann R K, Slayback D, Shabanov N V and Myneni R B 2001 Variations in northern vegetation activity inferred from satellite data of vegetation index during 1981 to 1999 *J. Geophys. Res.* 106 20069–83