Modeling the Effect of Variations in Snowpack-Disappearance Date on Surface-Energy Balance on the Alaskan North Slope

Authors: Ling, Feng, and Zhang, Tingjun

Source: Arctic, Antarctic, and Alpine Research, 37(4) : 483-489

Published By: Institute of Arctic and Alpine Research (INSTAAR), University of Colorado

URL: https://doi.org/10.1657/1523-0430(2005)037[0483:MTEOVI]2.0.CO;2

Arctic, Antarctic, and Alpine Research, Vol. 37, No. 4, 2005, pp. 483–489

Modeling the Effect of Variations in Snowpack-Disappearance Date on Surface-Energy Balance on the Alaskan North Slope

Feng Ling† and*
Tingjun Zhang‡*

*Institute of Plateau Meteorology, China Meteorological Administration, Chengdu, Sichuan 610072, People's Republic of China
†Department of Computer Science, Zhaoqing University, Zhaoqing, Guangdong 526061, People's Republic of China
‡Corresponding author. National Snow and Ice Data Center, Cooperative Institute for Research in Environmental Sciences, University of Colorado, Boulder, Colorado 80303-0449, U.S.A.
tzhang@kryos.colorado.edu

Abstract

Variations in the timing and duration of seasonal snow cover have a significant impact on local and regional atmospheric dynamics, hydrologic and geomorphic processes through the modification of the surface-radiation balance and the ultimate effect on near-surface-air temperature. This study investigated numerically the effect of variations in the date of seasonal snow-cover disappearance on the components of surface-energy balance on the Alaskan North Slope by using a heat-transfer model with phase change. The baseline inputs for the meteorological characteristics included mean daily air temperature, dew-point temperature, snow-cover depth, incident solar radiation, wind speed, and atmospheric pressure observed at Barrow, Alaska, from 1995 to 1998. Three simulation cases were conducted by using the measured snow data and by varying the snowpack-disappearance date by 10 days in 1998. The mean differences of the components of surface-energy balance caused by variations in snowpack disappearance were also quantified on an annual basis from July 1997 through June 1998. Results indicate that varying the snowpack-disappearance date by 10 days in spring can strongly affect the mean annual net solar radiation, sensible heat flux, and latent heat flux and can slightly affect the mean annual net longwave radiation and conductive heat flux.

Introduction

Seasonal snow cover significantly affects energy and mass exchanges between the ground surface and the atmosphere. The high surface albedo and emissivity of snow reduce the amount of absorbed solar energy and increase the amount of outgoing longwave radiation (e.g., Robock, 1980; Cohen and Rind, 1991; Robinson et al., 1993; Groisman et al., 1994a; Zhang et al., 1997a; Leathers and Luff, 1997; Stroeve and Nolin, 2002). The low thermal conductivity of snow insulates the ground, increasing the ground temperature during the cold season (e.g., Lachenbruch, 1959; Outcalt et al., 1975; Smith, 1975; Goodrich, 1982; Benson and Sturm, 1993; Zhang et al., 1996a; Sturm et al., 1997). In addition, snow cover represents a significant heat sink during the warming period of the seasonal cycle, owing to snow's high latent heat of fusion; snow cover thus provides a major source of thermal inertial within the climate system as it takes in and releases large quantities of energy with little or no fluctuation in temperature (Cohen and Rind, 1991; Barry et al., 1995; Strack et al., 2003). Studies have shown that the Earth's climate system is very sensitive to the large-scale distribution of snow cover (Walsh et al., 1982; Groisman et al., 1994a; Leathers and Robinson, 1997), and snow cover is a crucial issue for studies of climate change and for weather forecasting (Gustafsson et al., 2001).

Several studies have documented significant changes in the seasonal and interannual features of snow cover in the Northern Hemisphere over the past several decades. On the basis of snow-depth measurements, Foster (1989) reported that the date of spring snowmelt at Barrow, Alaska, occurred more than 2 wk earlier in the 1980s than in the 1950s. In a follow-on study, Foster et al. (1992) analyzed visible-band satellite snow-cover data for the 1976–1990 period in the arctic regions of Alaska, Canada, Scandinavia, and Siberia and observed consistent trends toward earlier spring snow-cover disappearance in all regions except Siberia. Ye (2001) noted that the length of the snow season over north-central and northwest Asia has increased by about four days per decade on the basis of historical records during 1937–1994. Dye (2002) examined the annual snow-cover cycle by using a 29-yr time series (1972–2000) of weekly visible-band satellite observations of the Northern Hemisphere snow cover; Dye demonstrated substantial interannual variability in the week of the first-observed snow cover in autumn and the week of the last-observed snow cover in spring, with a standard deviation of 0.7–0.9 and 0.8–1.1 wk.

Variations in snowpack-disappearance dates in spring not only influence local- and regional-scale heat- and water-exchange processes, but also affect the Earth's global heat budget and climate fluctuations (Robinson et al., 1986; Barnett et al., 1989; Baker et al., 1992). Jones and Briffa (1992) reported that over the past 100 yr, global surface-air temperature significantly increased. Century-long trends in surface-air temperature were especially prominent over the land area of the Northern Hemisphere during April and May. Groisman et al. (1994b) suggested that there is a strong positive feedback between spring snow cover and the radiative balance over northern extratropical land. The global warming that has occurred in the spring during the twentieth century is likely to have been significantly enhanced by corresponding changes in snow-cover extent. Thus, the transient effects of external climate forcing, such as greenhouse warming, may be especially prominent during spring in the Northern Hemisphere, where snow-cover variations substantially affect the radiative balance.

Snow covers the ground surface for as long as nine months of each year on the Alaskan North Slope. The date on which seasonal snow cover disappears ranges from late May to the middle of June (Kane et al., 1991; Zhang et al., 1997a). A recent study focusing on Barrow, Alaska, shows that the spring snowmelt has advanced by about eight days on average since the mid-1960s (Stone et al., 2002). Variations in the date of seasonal snow-cover disappearance can substantially affect macro- and microclimate conditions through the surface-energy balance. The date on which the seasonal snow cover melts has been regarded as a potential indicator of climate change in the Arctic (Foster, 1989; Curry and Ebert, 1992; Zhang et al., 1997b;

1523-0430/05 $7.00

Stone et al., 2002). A better understanding of the effect of variations in the snowpack-disappearance date on surface-energy balance is essential for the studies of climate dynamics and climate change in the Arctic. This is particularly important in view of the Global Circulation Model (GCM) prediction that the effect of global change will occur earlier and be greater in the Arctic than elsewhere (IPCC, 1990, 1996). Global warming will reduce the arctic snow and ice cover, which will result in a decrease in surface albedo. Lower albedo will increase solar absorption by the surface and promote further warming.

In the previous study (Ling and Zhang, 2003), we investigated the impact of changes in the timing and duration of seasonal snow cover on the thermal regime of the active layer and permafrost at Barrow, Alaska, by using a heat-transfer model based on an approach involving surface-energy balance (Ling and Zhang, 2004). We found that ground temperature is sensitive to variations in the timing and duration of seasonal snow cover. In the present study, we quantitatively estimate the effect of variations in the snowpack-disappearance date on the components of surface-energy balance by using the heat-transfer model and meteorological data from Barrow, Alaska.

Model Description and Data Sources

A one-dimensional heat-transfer model with phase change, combined with a surface-energy balance equation, was employed to investigate the effect of variations in the disappearance date of seasonal snow cover on surface-energy balance. The surface-energy balance equation was used to calculate surface heat-flux components and to estimate the upper-boundary temperature conditions for calculations of thermal conduction. The influence of unfrozen water on the thermal properties of frozen soils was included in the heat-transfer model. The effect of snow was also included in the model by extending the heat-conduction solution into the snow layer and computing the surface heat-balance components and the snow-surface temperature. A detailed description of the model development is given by Ling and Zhang (2004). Only a brief introduction, emphasizing the calculations for surface heat-flux components, is presented here.

The equation of energy fluxes at the surface can be expressed as

$$(1-\alpha)Q_{si} + Q_{li} + Q_{le} + Q_h + Q_e + Q_c = Q_m, \tag{1}$$

where α is the albedo of the surface, Q_{si} is the solar radiation reaching the surface of the Earth, Q_{li} is the incoming longwave radiation, Q_{le} is the emitted longwave radiation, Q_h is the turbulent exchange of sensible heat, Q_e is the turbulent exchange of latent heat, Q_c is the conduction heat flux through the snow cover or ground surface from below, and Q_m is the energy flux available for snow melting. All the energy-balance components have units of watts per square meter (W m^{-2}).

The incoming longwave radiation is given by the empirical description (Satterlund, 1979; Fleagle and Businger, 1980)

$$Q_{li} = 1.08\left(1-\exp\left(-(0.01e_a)^{\frac{T_a}{2016}}\right)\right)\sigma T_a^{\,4} \tag{2}$$

$$\log_{10}(e_a) = 11.40 - \frac{2353}{\mathrm{T_{dp}}}, \tag{3}$$

where T_a is mean daily air temperature (in kelvins, K), T_{dp} is daily dew-point temperature (in K), σ is the Stefan-Boltzmann constant (in W m^{-2} K^{-4}), and e_a is atmospheric water-vapor pressure (in pascals, Pa).

The emitted longwave radiation is given by

$$Q_{le} = -\varepsilon_s \sigma T_{s0}^{\,4}, \tag{4}$$

where ε_s is the surface emissivity and T_{s0} is ground surface or snow-surface temperature (in K).

The turbulent exchange of sensible and latent heat, Q_h and Q_e, the exchange coefficients for sensible heat and latent heat, D_h and D_e, and the stability function, ξ, are given by the following equations of Price and Dunne (1976):

$$Q_h = \rho_a C_P D_h \xi (T_a - T_{s0}) \tag{5}$$

$$Q_e = \rho_a L_s D_e \xi \left(0.622\frac{e_a - e_{s0}}{P_a}\right) \tag{6}$$

$$D_h = D_e = \frac{\kappa^2 U_z}{(\ln(z/z_0))^2} \tag{7}$$

$$\xi = \frac{1}{1+10_{R_i}}, \tag{8}$$

where ρ_a is the density of the air (in kg m^{-3}), assumed to be 1.275 kg m^{-3}, C_P is the specific heat of air (in J K^{-1} kg^{-1}), assumed to be 1004.0 J K^{-1} kg^{-1}, e_{s0} is the water-vapor pressure of the surface (in Pa), P_a is atmospheric pressure (in Pa), L_s is the latent heat of sublimation (in J kg^{-1}), κ is Von Karman's constant, U_z is the wind speed (in m s^{-1}) at reference height z (in m), and z_0 is the roughness length (in m). The Richardson number, R_i, is given by (Myrup, 1969)

$$R_i = \frac{gz(T_a - T_{s0})}{T_a U_z^{\,2}}, \tag{9}$$

where g is the gravitational acceleration (in m s^{-2}).

Heat conducted through the snow and ground is (Liston and Hall, 1995)

$$Q_c = -(T_{s0} - T_b)\left(\frac{z_s}{k_s} + \frac{z_g}{k_g}\right)^{-1}, \tag{10}$$

where T_b is the ground temperature at the bottom of top layer (in K), z_s, and z_g are the thicknesses of snow and the top layer of ground (in m), respectively, and k_s and k_g are the thermal conductivities of the snow and ground (in W m^{-1} K^{-1}), respectively.

The primary focus of this study is to quantify the effect of variations in the snow-cover disappearance date on surface-energy flux in the winter of 1997–1998. In order to minimize the effect of the assumed initial temperature distribution, the simulation was started at a model time of January 1995. The basic meteorological data used in this study include mean daily air temperature, dew-point temperature, snow-cover depth, wind speed, and atmospheric pressure (measured at the National Weather Service [NWS] station at Barrow, Alaska) and the incident solar radiation (measured at the NOAA Climate Monitoring and Diagnostics Laboratory [CMDL] at Barrow). These values are described by Ling and Zhang (2004).

The effective thermal conductivity, k_s (in W m^{-1} K^{-1}), and the volumetric heat capacity of snow, C_s (in J m^{-3} K^{-1}), were increased with increasing snow density, ρ_s (in kg m^{-3}) on the basis of the following empirical formulas (Goodrich, 1982):

$$k_s = 2.9\times10^{-6}\rho_s^{\,2} \tag{11}$$

$$C_s = 2.09\times10^{3}\rho_s. \tag{12}$$

The albedo of snow, α_s, decreases with increasing of snow density, ρ_s (Liston and Hall, 1995):

$$\alpha_s = \begin{cases} 1.0 - 0.247[0.16 + 110(\rho_s/1000)^4]^{1/2} & 50 \le \rho_s \le 450 \\ 0.6 - \rho_s/4600 & \rho_s > 450 \end{cases}. \tag{13}$$

Previous studies on the Alaskan North Slope have suggested a basis for dividing the snow season into stages (Kelly and Weaver, 1969; Maykut and Church, 1973; Weller and Holmgren, 1974; Sturm, 2000). At Barrow, Alaska, the snow season of 1997–1998 started on 22 September 1997 and ended with snow disappearance on 28 May 1998. The season was divided into four periods: formation, accumulation A, accumulation B, and melt, on the basis of previous studies and the validated model (Ling and Zhang, 2004). The 200 days of accumulation A and accumulation B periods were extended to 210 days by

TABLE 1

The four periods during the snow season of 1997–1998 for the three simulation cases conducted in this study.

	Snow season of 1997–1998			
Simulation case	Formation	Accumulation A	Accumulation B	Melt
SC0	22 Sept.–26 Oct. 1997	27 Oct.–15 Dec. 1997	16 Dec. 1997–14 May 1998	15 May–28 May 1998
SC1	22 Sept.–26 Oct. 1997	27 Oct.–13 Dec. 1997	14 Dec. 1997–04 May 1998	05 May–18 May 1998
SC2	22 Sept.–26 Oct. 1997	27 Oct.–17 Dec. 1997	18 Dec. 1997–24 May 1998	25 May–07 June 1998

linear interpolation, and were also shortened to 190 days by removing 10 days linearly. The snow depth for each interpolated day was derived by using a 6-day average snow depth of 3 days before and 3 days after the interpolated day. Three simulation cases were conducted (Table 1) by using the measured data (SC0), advancing the snowpack-disappearance date by 10 days in spring (SC1), and delaying the snowpack-disappearance date by 10 days in spring (SC2).

The snow density in each period in the snow season of 1997–1998 was chosen as follows (Ling and Zhang, 2004):

$$\rho_s = \begin{cases} 154 & \text{formation} \\ 256 & \text{accumulation } A \\ 366 & \text{accumulation } B \\ 385 & \text{melt.} \end{cases} \tag{14}$$

The albedo of the tundra surface (peat) at Barrow was chosen as 0.17 (Outcalt et al., 1975); the emissivity and roughness length of snow and the ground surfaces at different periods are summarized in Table 2.

The daily mean air temperature and dew-point temperature were assumed not to change with variations in the disappearance date of seasonal snow cover. This assumption is based on the fact that the energy available to melt snow at high latitudes is almost exclusively provided by radiation (Weller et al., 1972; Zhang et al., 1996b; Cline, 1997). Solar energy determines the period of possible snowmelt, whereas downwelling atmospheric longwave radiation modifies the timing and triggers the onset of snowmelt (Zhang et al., 2001). Studies have shown that snow melts because of radiative heating even when the air temperature is a few degrees below the freezing point (Koh and Jordan, 1995; Zhang et al., 2001). On the other hand, snow cover could be present when the surface temperature is $>0°C$, and under these circumstances, energy is available to either ripen or melt the snow cover (Outcalt et al., 1975; Liston and Hall, 1995).

Results and Discussion

The components of surface-energy balance in the snow season of 1997–1998 at Barrow, Alaska, have been simulated and discussed by Ling and Zhang (2004). This section examines the effect of variations in snowpack-disappearance date in the spring on the components of surface-energy balance during the snow season of 1997–1998. We define the energy fluxes toward the surface as positive.

EFFECT ON THE NET SOLAR RADIATION

The simulated differences of daily net solar radiation, Q_{ns}, between SC1 and SC0 and between SC2 and SC0 from 17 May through 10 June 1998 are presented in Figure 1. For a 10-day advance in snowpack-disappearance date, the corresponding net solar radiation flux increases, with a maximum value of 180 W m^{-2}. Net solar radiation flux decreases in the simulation with a 10-day delay in snowpack-disappearance date; the maximum decrease is 165 W m^{-2}. This is not surprising, as changing the ground surface from snow covered to snow free leads to decreasing the surface albedo. Alternatively, changing ground surface from snow free to snow covered results in an increased surface albedo. The high surface albedo of snow reduces the amount of net solar radiation by increasing upwelling solar radiation. Because the incident solar radiation on the Alaskan North Slope in spring is very high, any variation in snowpack-disappearance date has a significant impact on the net solar radiation. Stone et al. (2002) reported that the net radiative forcing can exceed 150 W m^{-2} on a daily basis immediately following the last day of snowmelt at Barrow, Alaska. The simulated results from our model agree well with their result.

EFFECT ON THE NET LONGWAVE RADIATION

From equation 4, surface temperature (snow-surface temperature when snowpack was present and ground-surface temperature when snowpack was absent), T_{s0}, is an important factor in determining the magnitude of net longwave radiation. To better understand the effect of variations in snowpack-disappearance date on the net longwave radiation, it is helpful to examine the response of the surface temperature to variations in snowpack-disappearance date.

Figure 2a is the simulated daily surface temperatures for SC0 (solid line), SC1 (dotted line), and SC2 (dashed line). Setting the snowpack-disappearance date 10 days early in spring results in an increased surface temperature, with a value of generally $<1.7°C$. This is due to the obvious differences in surface albedo and surface emissivity between snow and peat (Table 2). Delaying the snowpack-disappearance date by 10 days leads to a decrease in surface temperature, with a maximum decrease of 7.0°C. This change is primarily due to the significant heat-sink feature of snow cover, owing to its high latent heat of fusion. The calculated surface temperatures from baseline inputs were $>0°C$ during the 10 days from 29 May through 7 June 1998 (solid lines in Fig. 2a). Changes in the ground surface from snow free to snow covered leads to keeping the surface temperature at 0°C because of the existence of the snow cover, thus decreasing the surface temperature considerably.

The net longwave radiation, Q_{nl}, is affected by both the surface temperature and the surface emissivity. Advancing the snowpack-disappearance date by 10 days in spring results in the net longwave radiation becoming less negative (Fig. 2b). This is because the surface emissivity changed from 0.98 to 0.92 and the surface temperature increased (Fig. 2a). A decrease in surface emissivity leads to a decrease in the emitted longwave radiation, thus making the net longwave radiation

TABLE 2

Emissivity and roughness length for snow and peat surfaces at different periods.

Period	Emissivity	Roughness length (m)
Snow formation and accumulation	0.98[a]	0.005[a]
Snow melt	0.96[b]	0.015[c]
Snow free (peat surface)	0.92[d]	0.015[c]

[a] Liston and Hall (1995).
[b] Ling and Zhang (2004).
[c] Outcalt et al. (1975).
[d] Miller (1979).

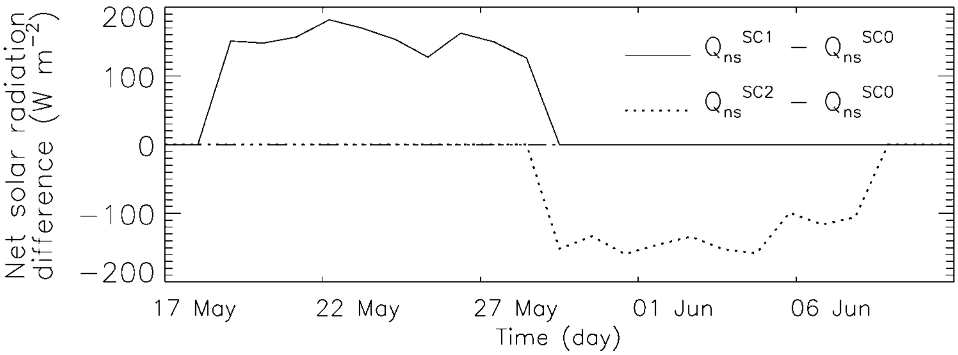

FIGURE 1. Simulated net solar radiation differences between SC1 and SC0 (solid line) and between SC2 and SC0 (dotted line) from 17 May through 10 June 1998. Energy fluxes toward the surface are defined to be positive.

less negative. An increase in surface temperature leads to an increase in the emitted longwave radiation, thus resulting in more negative net longwave radiation. The net effect of an early snowpack-disappearance date is that the net longwave radiation is less negative, because the surface-temperature increase is relatively small (Fig. 2a). Delaying the snowpack-disappearance date by 10 days in spring also results in less negative net longwave radiation (Fig. 2b). This is because the surface emissivity changed from 0.92 to 0.98 and the surface temperature decreased considerably (Fig. 2a). An increase in surface emissivity leads to an increase in emitted longwave radiation, thus making the net longwave radiation more negative. On the other hand, a decrease in surface temperature leads to a decrease in the emitted longwave radiation, thus resulting in less negative net longwave radiation. Because the surface temperature decrease is relatively large (generally greater than 4.0°C), the net effect is that the net longwave radiation is less negative.

EFFECT ON THE SENSIBLE AND LATENT HEAT FLUXES

On the basis of equation 5, the magnitude and variability of surface sensible heat flux, Q_e, is strongly affected by the difference between air temperature and surface temperature. Because of the variations in surface temperature caused by changes in the snowpack-disappearance date in spring (Fig. 2a), the surface sensible heat flux changes significantly. An early snowpack-disappearance date in spring results in a more negative sensible heat flux, whereas a delayed snowpack-disappearance date 10 days later leads to a less negative sensible heat flux (Fig. 3a).

Figure 3b shows the simulated latent heat fluxes, Q_h, for SC0, SC1, and SC2 from 17 May through 10 June 1998. Setting the snowpack-disappearance date 10 days early in spring results in a negative latent heat flux and produces a maximum heat-flux variation of nearly 150 W m^{-2}. Setting the snowpack-disappearance date 10 days later in spring leads to a positive latent heat flux, with a maximum variation >200 W m^{-2}. This is true because advancing the snowpack-disappearance date by 10 days increases the surface temperature (Fig. 2a), leading to increased evaporation and reflecting the progressive thawing process of the active layer. Alternatively, delaying the snowpack-disappearance date by 10 days leads to a decreased surface temperature; thus the latent heat flux becomes positive, reflecting the decrease in evaporation.

EFFECT ON THE CONDUCTIVE HEAT FLUX

Conductive heat flux through the ground surface and snow cover, Q_c, is related to the difference between the surface temperature, T_{s0},

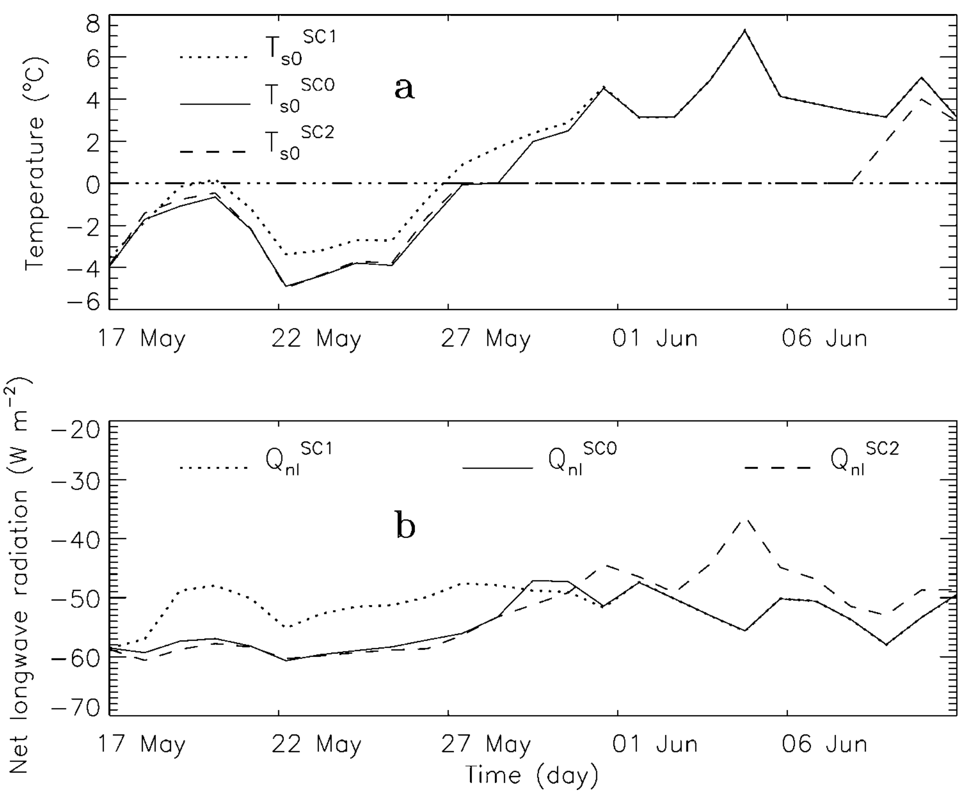

FIGURE 2. Comparisons of simulated (a) daily surface temperatures for SC0 (solid line), SC1 (dotted line), and SC2 (dashed line) and (b) net longwave radiation fluxes for SC0 (solid line), SC1 (dotted line), and SC2 (dashed line) from 17 May through 10 June 1998.

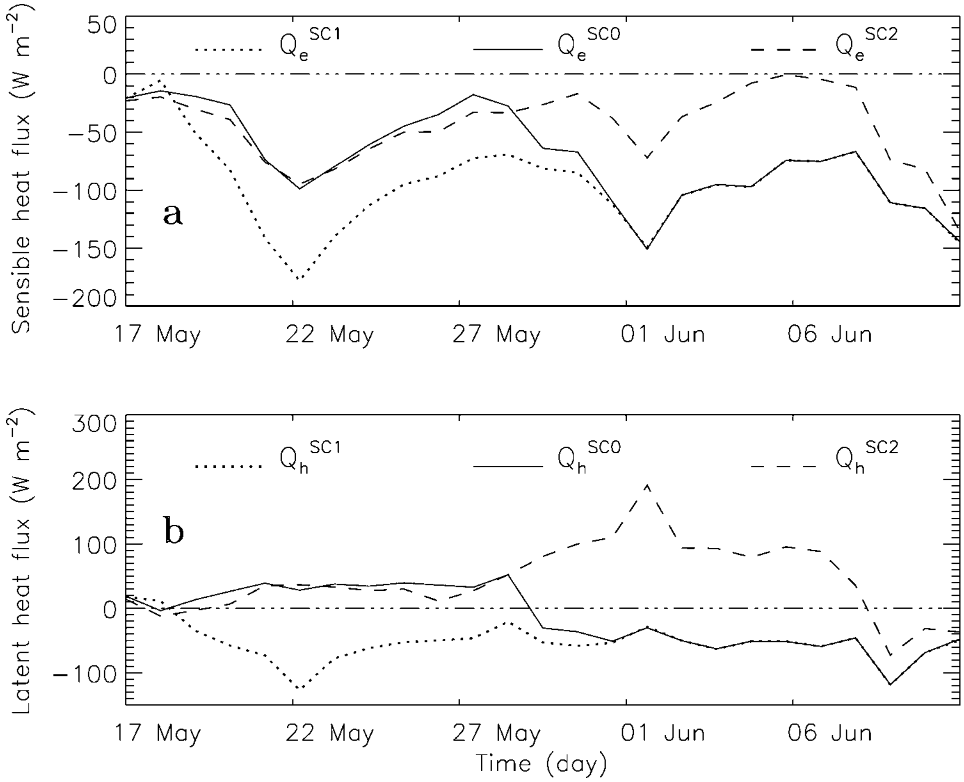

FIGURE 3. Comparisons of simulated (a) sensible heat fluxes for SC0 (solid line), SC1 (dotted line), and SC2 (dashed line) and (b) latent heat fluxes for SC0 (solid line), SC1 (dotted line), and SC2 (dashed line) from 17 May through 10 June 1998.

and the ground temperature at the bottom of the top layer, T_b. This study sets the bottom of the top node layer at a depth of 0.03 m. Figure 4a presents the simulated differences between the surface temperature and the ground temperature at a depth of 0.03 m for SC0, SC1, and SC2.

Setting the snowpack-disappearance date 10 days early in spring slightly decreases the difference between the surface temperature and the ground temperature at a depth of 0.03 m. Delaying the snowpack-disappearance date 10 days in spring also results in a decreasing difference between the surface temperature and the ground temperature at a depth of 0.03 m. The fact that both simulations result in a decreased difference in the two temperatures can be explained as follows: The ground temperature at a depth of 0.03 m is, in general, higher than the snow-surface temperature when snow cover is present

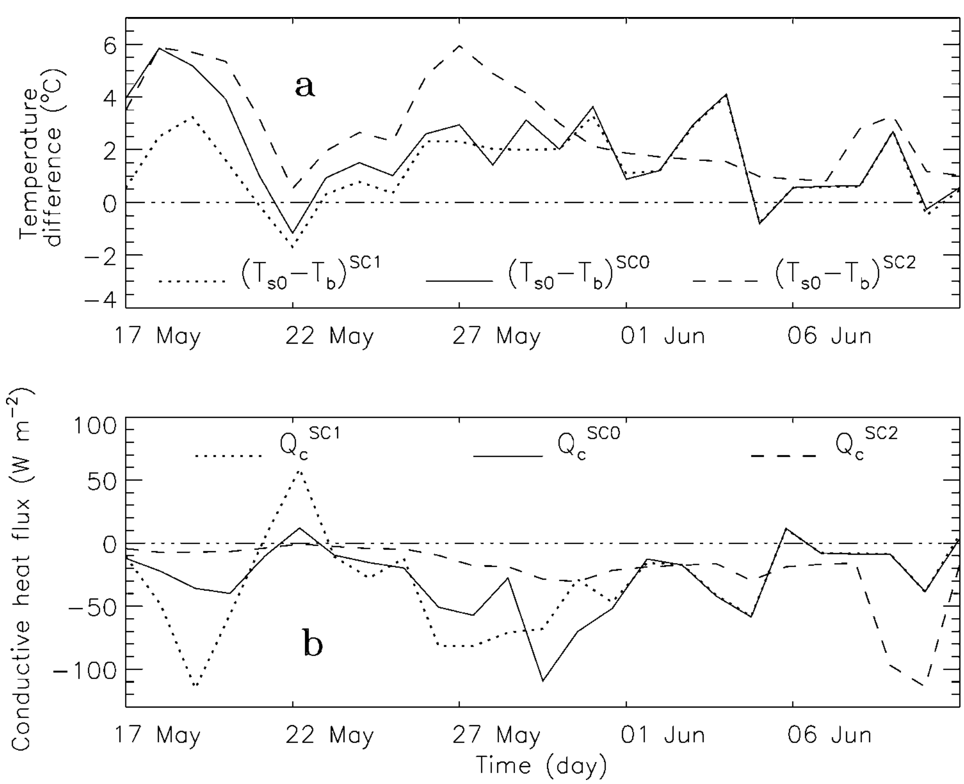

FIGURE 4. Simulated (a) temperature differences between surface temperature and ground temperature at a depth of 0.03 m for SC0 (solid line), SC1 (dotted line), and SC2 (dashed line) and (b) conductive heat fluxes for SC0 (solid line), SC1 (dotted line), and SC2 (dashed line) from 17 May through 10 June 1998.

TABLE 3

Comparison of the mean annual differences of simulated components of surface-energy balance (W m^{-2}) between SC1 and SC0 and between SC2 and SC0 from July 1997 through June 1998.

	Mean annual difference	
Simulated components	Between SC1 and SC0	Between SC2 and SC0
Net solar radiation	4.18	−3.71
Net long-wave radiation	0.24	0.15
Sensible heat flux	−2.43	4.09
Latent heat flux	−1.46	1.94
Conductive heat flux	−0.53	0.17

and varies relatively slowly with changes in the snow-surface temperature owing to the insulation effect of snow cover. Advancing the snowpack-disappearance date by 10 days increases the surface temperature, and therefore, slightly decreases the difference between the surface temperature and ground temperature at a depth of 0.03 m, because the thermal conductivity of peat is relatively higher than that of snow. Consequently, the temperature difference for SC1 is less than that for SC0. After snow disappeared on May 29, the simulated ground surface temperature from SC0 was $>0°C$ (Fig. 2a). Delaying the snowpack-disappearance date by 10 days results in a constant surface temperature of 0°C. Snow was in an isothermal state at 0°C, and the difference between the snow-surface temperature and the ground temperature at a depth of 0.03 m for SC2 decreased slowly with time during the 10 days from May 29 to June 7, reflecting the snow-surface temperature of 0°C. Consequently, the difference between the surface temperature and the ground temperature at a depth of 0.03 m for SC2 slightly decreases with time.

Figure 4a also indicates that the simulated temperature differences between the surface temperature and the ground temperature at a depth of 0.03 for SC0, SC1, and SC2 are different even prior to changing the snowpack-disappearance dates. The reason is that the snow albedos in the accumulation B and melt periods (Table 1) have obvious differences caused by variations in snow density (eq. 13).

Besides the difference between surface temperature and ground temperature at the bottom of the top layer, conductive heat flux also strongly depends on the thermal conductivity of the mixture of snow and peat. Although advancing the snowpack-disappearance date by 10 days decreases the difference between the surface temperature and the ground temperature at a depth of 0.03 m (Fig. 4a), the magnitude of conductive heat flux increases (Fig. 4b). This is because advancing the snowpack-disappearance date results in an increased effective thermal conductivity, because the thermal conductivity of peat is higher than that of snow. Delaying the snowpack-disappearance date by 10 days leads to a small conductive heat flux (Fig. 4b), reflecting the small and slowly decreased difference between surface temperature and ground temperature at a depth of 0.03 m as well as the decrease in the effective thermal conductivity of snow and peat.

Variations in the net surface radiation budget result in a redistribution of energy in complicated ways that involve ground storage, sensible and latent heat exchanges between the surface and atmosphere, and advective processes that can distribute the radiation gain to other regions. Maykut and Church (1973) pointed out that the most significant factor influencing the magnitude of the yearly net radiation total is the date when snowmelt is completed. To further address the effect of variations in snowpack-disappearance date on the local climate system, Table 3 quantified the mean differences of the calculated components of surface-energy balance between simulation cases SC1 and SC0 and between SC2 and SC0 on an annual basis from July 1997 through June 1998. Varying the snowpack-disappearance date by 10 days in spring can strongly affect the mean annual net solar radiation, sensible heat flux, and latent heat flux and can slightly affect the mean annual net longwave radiation and conductive heat flux.

Summary and Conclusions

A surface-energy balance equation combined with a one-dimensional heat-transfer model with phase change was used to quantify the effect of variations in snowpack-disappearance date in spring on components of surface-energy balance. The model was driven with observed mean daily meteorological data collected at Barrow, Alaska. Three simulation cases were conducted by using the measured snow data and by changing the snowpack-disappearance date by 10 days before and 10 days after the actual date in spring in 1998.

Advancing the snowpack-disappearance date by 10 days in spring resulted in a net solar radiation increase and led to a mean annual increase of 4.18 W m^{-2}. Net longwave radiation became less negative; there was a mean annual increase of 0.24 W m^{-2}. The sensible heat flux became more negative; there was a mean annual decrease of 2.43 W m^{-2}, and the latent heat flux changed from positive to negative, which resulted in a mean annual decrease of 1.46 W m^{-2}. The conductive heat flux decreased, which led to a mean annual decrease of 0.53 W m^{-2}.

Delaying the snowpack-disappearance date by 10 days in spring led to a net solar radiation decrease and resulted in a mean annual decrease of 3.71 W m^{-2}. Net longwave radiation became less negative and showed a mean annual increase of 0.15 W m^{-2}. The sensible heat flux became more negative, which resulted in a mean annual decrease of 4.09 W m^{-2}. The latent heat flux changed from positive to negative and showed a mean annual decrease of 1.94 W m^{-2}. The conductive heat flux decreased, which led to a mean annual decrease of 0.17 W m^{-2}.

Acknowledgments

We would like to express our gratitude to the anonymous reviewers for their constructive and helpful comments. We also thank Lyne Yohe and Mary C. Eberle who kindly edited the manuscript. This work was in part supported by the Guangdong Natural Science Foundation (grant no. 04011600) of the People's Republic of China, the Scientific Research Foundation for the Returned Overseas Chinese Scholars, the U.S. National Science Foundation through the NSF OPP-0352910 and OPP-0229766, and the International Arctic Research Center, University of Alaska–Fairbanks, under the auspices of the NSF cooperative agreement number OPP-0327664. Any opinions, findings, and conclusions or recommendations expressed in this article are those of the authors and do not necessarily reflect the views of the funding institutes and agencies.

References Cited

Baker, D. L., Ruschy, R. H., Skahhs, and Wall, D. B., 1992: Air temperature and radiation depressions associated with a snow cover. *Journal of Applied Meteorology,* 31: 247–254.

Barnett, T. P., Dumenil, L., Schlese, U., Roeckner, E., and Latif, M., 1989: The effect of Eurasian snow cover on regional and global climate variations. *Journal of Atmospheric Sciences,* 46: 661–685.

Barry, R. G., Fallot, J.-M., and Armstrong, R. L., 1995: Twentieth-century variability in snow cover conditions and approaches to detecting and monitoring changes: Status and prospects. *Progress in Physical Geography,* 19: 520–532.

Benson, C. S., and Sturm, M., 1993: Structure and wind transport of seasonal snow on the arctic slope of Alaska. *Annuals of Glaciology,* 18: 261–267.

Cline, D. W., 1997: Snow surface energy exchanges and snowmelt at a continental, midlatitude Alpine site. *Water Resources Research,* 33: 689–701.

Cohen, J., and Rind, D., 1991: The effect of snow cover on the climate. *Journal of Climate,* 4: 689–706.

Curry, J. A., and Ebert, E. E., 1992: Annual cycle of radiation fluxes over the Arctic Ocean: Sensitivity to cloud optical properties. *Journal of Climate,* 5: 1267–1280.

Dye, D. G., 2002: Variability and trends in the annual snow-cover cycle in Northern Hemisphere land areas, 1972–2000. *Hydrological Processes,* 16: 3065–3077.

Fleagle, R. G., and Businger, J. A., 1980: *An introduction to atmospheric physics.* New York: Academic Press, 432 pp.

Foster, J. L., 1989: The significance of the date of snow disappearance on the arctic tundra as a possible indicator of climate change. *Arctic and Alpine Research,* 21: 60–70.

Foster, J. L., Winchester, J. W., and Dutton, W. G., 1992: The date of snow disappearance on the arctic tundra as determined from satellite, meteorological station and radiometric in situ observations. *IEEE Transactions on Geoscience and Remote Sensing,* 30: 793–798.

Goodrich, L. E., 1982: The influence of snow cover on the ground thermal regime. *Canadian Geotechnical Journal,* 19: 421–432.

Groisman, P. Y., Karl, T. R., Knight, R. W., and Stenclukov, G. L., 1994a: Changes of snow cover, temperature, and radiative heat balance over the Northern Hemisphere. *Journal of Climate,* 7: 1633–1656.

Groisman, P. Y., Karl, T. R., and Knight, R. W., 1994b: Observed impact of snow cover on the heat balance and the rise of the continental spring temperatures. *Science,* 263: 198–200.

Gustafsson, D., Stahli, M., and Jansson, P. E., 2001: The surface energy balance of a snow cover: Comparing measurements to two different simulation models. *Theoretical and Applied Climatology,* 70: 81–96.

Intergovernmental Panel on Climate Change (IPCC), 1990: *Climate Change: The IPCC Scientific Assessment.* New York: Cambridge University Press, 364 pp.

Intergovernmental Panel on Climate Change (IPCC), 1996: *Climate Change 1995: The Science of Climate Change.* New York: Cambridge University Press, 572 pp.

Jones, P. D., and Briffa, K. R., 1992: Global surface air temperature variations during the twentieth century. Part I: Spatial, temporal and seasonal details. *The Holocene,* 2: 165–179.

Kane, D. L., Hinzman, L. D., and Zarling, J. P., 1991: Thermal response of the active layer to climate warming in a permafrost environment. *Cold Regions Science and Technology,* 19: 111–122.

Kelly, J. J., Jr., and Weaver, D. F., 1969: Physical processes at the surface of the arctic tundra. *Arctic,* 22: 425–437.

Koh, G., and Jordan, R., 1995: Sub-surface melting in a seasonal snow cover. *Journal of Glaciology,* 41: 474–482.

Lachenbruch, A. H., 1959: *Periodic heat flow in a stratified medium with application to permafrost problems.* Washington, D.C.: U.S. Geological Survey Bulletin 1083-A, 36 pp.

Leathers, D. J., and Luff, B. L., 1997: Characteristics of snow cover duration across the northeast United States of America. *International Journal of Climatology,* 17: 1535–1547.

Leathers, D. J., and Robinson, D. A., 1997: Abrupt changes in the seasonal cycle of North America snow cover. *Journal of Climate,* 10: 2569–2585.

Ling, F., and Zhang, T., 2003: Impact of the timing and duration of seasonal snow cover on the active layer and permafrost in the Alaskan Arctic. *Permafrost and Periglacial Processes,* 14: 141–150.

Ling, F., and Zhang, T., 2004: A numerical model for surface energy balance and thermal regime of the active layer and permafrost containing unfrozen water. *Cold Regions Science and Technology,* 38: 1–15.

Liston, G. E., and Hall, D. K., 1995: An energy-balance model of lake-ice evolution. *Journal of Glaciology,* 41: 373–382.

Maykut, G. A., and Church, P. E., 1973: Radiation climate of Barrow, Alaska: 1962–1966. *Journal of Applied Meteorology,* 12: 620–628.

Miller, T. W., 1979: The surface heat balance in simulations of permafrost behavior. *Journal of Energy Resources Technology,* 101: 240–250.

Myrup, L. G., 1969: A numerical model of the urban heat island. *Journal of Applied Meteorology,* 8: 908–918.

Outcalt, S. L., Goodwin, C., Weller, G., and Brown, J., 1975: Computer simulation of the snowmelt and soil thermal regime at Barrow, Alaska. *Water Resources Research,* 11: 709–715.

Price, A. D., and Dunne, T., 1976: Energy balance computations of snow melt in a sub-arctic area. *Water Resources Research,* 12: 686–689.

Robinson, D. A., Scharfen, G., Serreze, M. C., Kukla, G., and Barry, R. G., 1986: Snow melt and surface albedo in the Arctic Basin. *Geophysical Research Letters,* 13: 945–948.

Robinson, D. A., Dewey, K. F., and Heim, R. R., Jr., 1993: Global snow monitoring: An update. *Bulletin of the American Meteorological Society,* 74: 1557–1560.

Robock, A., 1980: The seasonal cycle of snow cover, sea ice and surface albedo. *Monthly Weather Review,* 108: 267–285.

Satterlund, D. R., 1979: An improved equation for estimating long-wave radiation from the atmosphere. *Water Resources Research,* 15: 1649–1650.

Smith, M. W., 1975: Microclimatic influence on ground temperatures and permafrost distribution, Mackenzie Delta, Northwest Territories. *Canadian Journal of Earth Sciences,* 12: 1421–1438.

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. *Journal of Geophysical Research,* 107(D10): ACL10 1–15.

Strack, J. L., Pielke SR, R. A., and Adegoke, J., 2003: Sensitivity of model-generated daytime surface heat fluxes over snow to land-cover changes. *Journal of Hydrometeorology,* 4: 24–42.

Stroeve, J. C., and Nolin, A. W., 2002: New methods to infer snow albedo from the MISR instrument with applications to the Greenland ice sheet. *IEEE Transactions on Geoscience and Remote Sensing,* 40: 1616–1625.

Sturm, M., 2000: Arctic winter and snow cover: Five distinct stages with implications for living things. *Eos (Transactions, American Geophysical Union),* 81: F232.

Sturm, M., Holmgren, J., Konig, M., and Morris, K., 1997: The thermal conductivity of seasonal snow. *Journal of Glaciology,* 43: 26–41.

Walsh, J. E., Tucek, D. R., and Peterson, M. R., 1982: Seasonal snow cover and short-term climatic fluctuations over the United States. *Monthly Weather Review,* 110: 1474–1485.

Weller, G., and Holmgren, B., 1974: The microclimates of the arctic tundra. *Journal of Applied Meteorology,* 13: 854–862.

Weller, G. S., Cubley, S., Parker, S., Trabant, D., and Benson, C., 1972: The tundra microclimate during snow-melt at Barrow, Alaska. *Arctic,* 25: 291–300.

Ye, H., 2001: Increases in snow season length due to earlier first snow and later last snow dates over north central and northwest Asia during 1937–94. *Geophysical Research Letters,* 28: 551–554.

Zhang, T., Osterkamp, T. E., and Stamnes, K., 1996a: Influence of the depth hoar layer of the seasonal snow cover on the ground thermal regime. *Water Resources Research,* 32(7): 2075–2086.

Zhang, T., Stamnes, K., and Bowling, S. A., 1996b: Impact of clouds on surface radiative fluxes and snowmelt in the Arctic and Subarctic. *Journal of Climate,* 9: 2110–2123.

Zhang, T., Stamnes, K., and Bowling, S. A., 1997a: Impact of the atmosphere on surface radiative fluxes and snowmelt in the Arctic and Subarctic. *Journal of Geophysical Research,* 102(D4): 4287–4302.

Zhang, T., Osterkamp, T. E., and Stamnes, K., 1997b: Effect of climate on the active layer and permafrost on the North Slope of Alaska, U.S.A. *Permafrost and Periglacial Processes,* 8: 45–67.

Zhang, T., Stamnes, K., and Bowling, S. A., 2001: Impact of the atmospheric thickness on the atmospheric downwelling long-wave radiation and snowmelt under clear-sky conditions in the Arctic and Subarctic. *Journal of Climate,* 14: 920–939.

Revised ms submitted February 2005