INFLUENCE OF THE SEASONAL SNOW COVER ON THE GROUND THERMAL REGIME: AN OVERVIEW

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[1] The presence of seasonal snow cover during the cold season of the annual air temperature cycle has significant influence on the ground thermal regime in cold regions. Snow has high albedo and emissivity that cool the snow surface, high absorptivity that tends to warm the snow surface, low thermal conductivity so that a snow layer acts as an insulator, and high latent heat due to snowmelt that is a heat sink. The overall impact of snow cover on the ground thermal regime depends on the timing, duration, accumulation, and melting processes of seasonal snow cover; density, structure, and thickness of seasonal snow cover; and interactions of snow cover with micrometeorological conditions, local microrelief, vegetation, and the geographical locations. Over different timescales either the cooling or warming impact of seasonal snow cover may dominate. In the continuous permafrost regions, impact of seasonal snow cover can result in an increase of the mean annual ground and permafrost surface temperature by several degrees, whereas in discontinuous and sporadic permafrost regions the absence of seasonal snow cover may be a key factor for permafrost development. In seasonally frozen ground regions, snow cover can substantially reduce the seasonal freezing depth. However, the influence of seasonal snow cover on seasonally frozen ground has received relatively little attention, and further study is needed. Ground surface temperatures, reconstructed from deep borehole temperature gradients, have increased by up to 4°C in the past centuries and have been widely used as evidence of paleoclimate change. However, changes in air temperature alone cannot account for the changes in ground temperatures. Changes in seasonal snow conditions might have significantly contributed to the ground surface temperature increase. The influence of seasonal snow cover on soil temperature, soil freezing and thawing processes, and permafrost has considerable impact on carbon exchange between the atmosphere and the ground and on the hydrological cycle in cold regions/cold seasons.

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1. INTRODUCTION

[2] The importance of the seasonal snow cover to ground thermal conditions is widely recognized. The effect of the seasonal snow cover on the ground thermal regime can be explained with reference to the pattern of seasonal temperature variation. Snow is an insulator compared to other natural materials and is a leading factor in protecting the ground from heat loss in winter. Although the net effect of the seasonal snow cover on the ground thermal regime is determined by many factors, in general, the seasonal snow cover tends to result in relatively higher mean annual ground temperatures, especially at high latitudes where stable snow cover lasts from a few weeks to several months. *Gold* [1963] concluded that snow cover was the principal reason that mean annual ground temperature can be many degrees warmer than the mean annual air temperature in cold regions. Snow amelioration measures have been widely used to mitigate ground freezing in agriculture, as well as in the open pit mining industry [*Klyukin*, 1963; *Nicholson*, 1978a, 1978b]. Where air temperature is close to 0°C, seasonal snow cover can be responsible for the absence of permafrost in certain locations [*Smith*, 1975; *Tong et al.*, 1983; *Zhang et al.*, 1985].

2. PHYSICAL PRINCIPLES

[3] Seasonal snow cover, which exists between the atmosphere and the ground surface during the cold seasons of the annual temperature cycle, is one of the primary factors influencing the ground thermal regime in cold regions. The influence of the seasonal snow cover on the ground thermal regime can be explained by the following factors.

2.1. High Surface Albedo

[4] The surface albedo is defined as the ratio of the all-wavelength solar radiation reflected by the surface to that incident upon it [*Henderson-Sellers and Hughes*, 1982; *Barry et al.*, 1984; *Barry*, 1996]. The snow surface spectrally integrated albedo, computed for the solar wavelengths (λ = 0.3–3.0 μm approximately), is a function of solar

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zenith angle, atmospheric properties, cloudiness, and surface characteristics. The optical properties of a snow cover that determine its albedo are the coefficients of spectral absorption and of volume scattering and the associated phase function of single scattering [*Maykut et al.*, 1992]. The respective coefficients characterize the fractional absorption and scattering from a light beam with a small-volume element, whereas the phase function describes the angular distribution of the scattering by this element. The surface albedo of a snow cover is influenced by three sets of factors: (1) the snow cover characteristics (grain size and shape, surface roughness, liquid water content, and any impurities); (2) the solar zenith angle and the cloud conditions; and (3) the albedo of the underlying surface up to some limiting threshold masking depth. *Barry* [1996] provided a detailed review on the factors determining the albedo of snow cover.

[5] Snow surface albedo ranges from less than 0.60 for wet and melting snow to greater than 0.85 for fresh snow. Snow albedo can be greater than 0.90 under cloudy sky conditions [*Wendler and Kelly*, 1988; *Zhang et al.*, 1996a]. The high snow surface albedo leads to a reduction in the absorbed solar energy and lowering of snow surface temperature. In general, the albedo of fresh snow surface in autumn is much higher than that for an old and melting snow surface in spring.

[6] Because of the changes in the solar elevation and the global radiation (the sum of the downwelling direct and diffuse solar radiation) with time and geographical location the albedo effect on the surface radiation balance, thus the surface temperature, also changes spatially and temporally. The solar elevation during the period of snow cover establishment in autumn is much lower than that during the period of snowmelt in spring, especially in the Arctic and sub-Arctic. Therefore the global radiation reaching the snow surface is much greater in spring than in autumn. The net albedo effect on the surface radiation balance, and hence the snow surface temperature, is greater in spring than in autumn even though the snow surface albedo is lower in spring than in autumn. For example, in northern Alaska, establishment of the stable snow cover generally occurs in late September and early October, while snowmelt usually happens in late May and early June [*Zhang et al.*, 1996a]. The global radiation varies from 40 to 80 W m^{-2} in September and October and from 240 to 280 W m^{-2} in May and June in Alaska (Figure 1) [see also *Serreze et al.*, 1998; *Stone et al.*, 1996, 2002]. If the albedo is 0.80 for fresh snow in autumn and 0.60 for melting snow in spring, the absorbed solar radiation will vary from only 8 to 16 W m^{-2} in autumn and from 96 to 112 W m^{-2} in spring. The net albedo effect on the surface radiation balance is about 8 to 15 times greater in spring than in autumn in northern Alaska. However, snowmelt at high latitudes is a rapid process that usually is completed within a week or 10 days [*Kane et al.*, 1991; *Zhang et al.*, 1996b; *Stone et al.*, 2002]. The overall snow albedo effect during the snowmelt period in spring on seasonal and annual timescales may be limited. The albedo effect on the snow surface temperature is also greater at lower and middle latitudes than at high latitudes because of the higher solar elevation. The cooling effect of snow albedo on snow surface is relatively limited during cold winter months, because of the lower solar elevation in winter, especially at high latitudes.

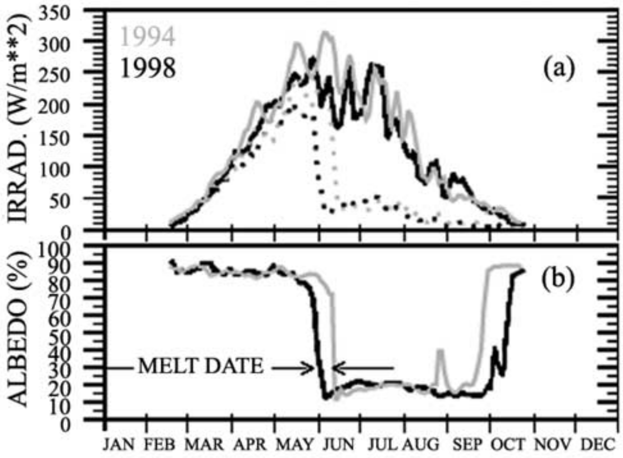

Figure 1. Seasonal cycles of (a) downwelling shortwave downward radiative flux (SD) (solid curves) and upwelling shortwave upward radiative flux (SU) (dashed curves) solar irradiance at NOAA/Climate Monitoring and Diagnostics Laboratory Barrow Observatory, representing late (1994) (shaded curves) and early (1998) (solid curves) spring melt seasons and (b) derived albedo (SU/SD) for the same 2 years [*Stone et al.*, 2002].

2.2. Emissivity and Absorptivity

[7] Over the thermal infrared part of the radiative spectrum, snow acts almost as a "blackbody." Emissivity of snow varies from 0.96 to 0.99, with an average of about 0.98. This value is generally higher than any other land surface emissivities. For example, the higher emissivity of a snow surface causes an increase in the outgoing longwave radiation, thus cooling the snow surface. For example, if the same amount of longwave radiation is released from a snow surface and a bare soil surface (no snow) with average emissivity of 0.98 and 0.92, respectively, the snow surface temperature is about 3.6°C to 4.4°C lower than the bare soil surface temperature.

[8] On the other hand, higher emissivity on the snow surface also implies higher absorptivity of snow surface as defined by Kirchhoff's law [*Wallace and Hobbs*, 1977]. Higher absorptivity means that a snow surface would absorb more energy than other types of surface conditions. This is particularly important during cloudy days in the Arctic. *Zhang et al.* [1996a] demonstrated that clouds could change the date of onset of snowmelt by as much as a month in the Arctic and sub-Arctic, mainly because of the enhanced atmospheric downwelling longwave radiation and near-blackbody absorption of snow surface.

[9] In general, for a cold, dry, and clear-sky condition, higher emissivity of snow surface may result in surface cooling and very often the development of a low-level temperature inversion [*Serreze et al.*, 1992]. For a relatively

wet atmosphere with cloudy sky, atmospheric downwelling longwave radiation is strong and the snow surface may absorb more energy because of its higher absorptivity, resulting in higher surface temperature. The overall effect of snow emissivity and absorptivity on snow surface temperature largely depends on overlying atmospheric conditions.

2.3. Low Thermal Conductivity

[10] Dry snow primarily consists of air and snowflakes or ice crystals. Thermal conductivity of pure ice at 0°C is about 2.24 W m^{-1} K^{-1} with a density of about 917 kg m^{-3}. Thermal conductivity of air is about 0.025 W m^{-1} K^{-1}, with a density of about 1.2 kg m^{-3}. Because a large fraction of the snow layer is filled with air, snow has an extremely low thermal conductivity. Thermal conductivity of the snow layer is primarily a function of snow density and ranges from less than 0.10 W m^{-1} K^{-1} for less dense fresh snow to greater than 0.5 W m^{-1} K^{-1} for more dense ripened snow, approximately 5 to 20 times lower than that of mineral soils.

[11] Because of its extremely low thermal conductivity, seasonal snow cover acts as an excellent insulator between the atmosphere and the ground surface. The soil surface beneath the snow cover can have either a lower or a much higher temperature than the snow surface and the air temperature, depending upon the timing, duration, and thickness of the seasonal snow cover and the air temperature history. The snow insulating effect prevents the soils from cooling. The insulation effect plays a major role in the influence of snow cover on the ground thermal regime as will be discussed in sections 3 and 4.

2.4. Latent Heat of Fusion

[12] The heat required for merely warming or cooling, that is, the heat capacities of the mineral or organic soil material, water, and ice, is relatively small by comparison with the quantity of latent heat of fusion. For example, the heat capacity for pure ice is about 2.1 J g^{-1} K^{-1}, while the latent heat of fusion is about 334 J g^{-1}. In other words, to warm 1 g of ice 1°C involves the addition of 2.1 J, while 334 J g^{-1} must be added to melt it. Therefore snowmelt is an energy sink because of the latent heat of fusion. Initially, melting snow keeps the ground surface at 0°C even though the air temperature could be well above 0°C. This leads to a certain cooling of the ground surface and lowering of the mean annual ground surface temperature [see *Zhang et al.*, 1997]. However, as snowmelt progresses, snow temperature increases rapidly and reaches the 0°C isotherm within a couple of days [*Dingman et al.*, 1980] because of the refreeze of meltwater and the release of latent heat to warm the entire snow layer, thus warming the soil surface. The overall impact of latent heat on surface temperature is complicated, and further study is needed. However, because snowmelt generally occurs very rapidly, usually within 1 to 2 weeks at high latitudes, the latent heat affects soil temperature for a very limited period of time.

[13] Other processes within the snow cover, such as nonconductive heat transfer, water vapor transfer, snow metamorphism, and the densification process, may have a different effect on the energy exchange between the atmosphere and the ground surface. Overall, high albedo and thermal emissivity of the seasonal snow cover tend to cool the snow surface and therefore the entire snowpack and soils underneath. According to meteorological data the mean winter temperature of a snow surface may be 0.5°C to 2.0°C lower than the mean winter air temperature [*Yershov*, 1998]. Latent heat due to snowmelt delays the soil surface warming and thus relatively cools the soils. Refreeze of meltwater within the snow cover releases latent heat and increases snow temperature and thus underlying ground surface and soil temperature. This leads to the insulation effect and higher absorptivity (especially under cloudy sky conditions) of the seasonal snow cover relatively warming the soil surface and underlying soils. The net effect of the seasonal snow cover on the ground thermal regime and its magnitude depends upon the timing, duration, accumulation, and melting processes of the seasonal snow cover; and the thickness, density, and structure of the seasonal snow cover; and interactions of micrometeorological conditions, local microrelief, vegetation, and the geographical locations.

3. IMPACT OF THE SEASONAL SNOW COVER ON GROUND TEMPERATURES

[14] Snow cover parameters, such as its timing, duration, thickness, density, and structure, change spatially and temporally. In this section, after a brief description of the methodology of air and soil temperature measurements, the impact of changes in these snow cover parameters and properties on the ground thermal regime will be discussed.

3.1. Methods of Soil Temperature Measurements and Uncertainties

[15] Temperature sensors for soil temperature measurements include mechanical devices (e.g., liquid-in-glass thermometers), resistance devices (e.g., thermistors), and voltage devices (e.g., thermocouples). Thermistors have relatively high measurement sensitivity and accuracy of up to <0.01°C and have become the most widely used sensors for precision temperature measurements in the permafrost study community during the last 2 decades. Sensitivity and accuracy of thermocouples fall in between those of the liquid-in-glass thermometers and the thermistors [*Osterkamp*, 1985]. Unless specified, both air and soil temperatures were measured using thermistors.

[16] Soil temperatures were measured from the ground surface to several hundred meters at different time intervals and with different instrumentation. For depths less than 1.0 m at various sites across Alaska, soil temperatures were measured by a string of thermistors every 4 hours, and the daily mean value was recorded on a computerized device. The thermistor string consists of nine individually calibrated thermistors mounted securely in a round plastic rod: Three of them were placed in the active layer, three were placed at depths bracketing the permafrost table, and three were placed in permafrost. The maximum and minimum ground

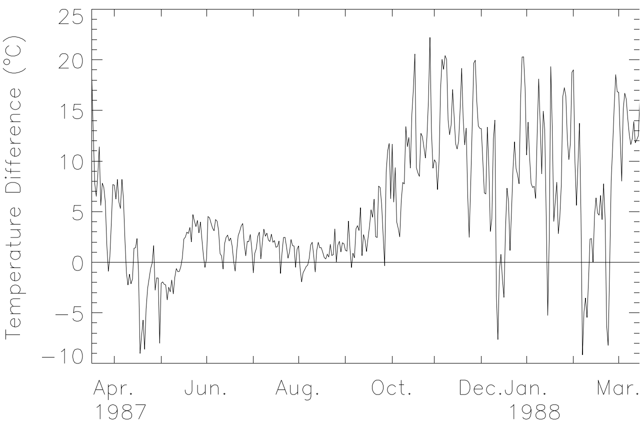

Figure 2. Temperature difference between daily mean ground surface and air temperatures at Franklin Bluffs, North Slope of Alaska, for the period from mid-March 1987 through mid-March 1988 [*Zhang et al.*, 1997]. The solid horizontal line is the 0°C reference line. Copyright John Wiley and Sons Ltd. Reproduced with permission.

surface and air temperatures (measured at 1.5 m above the ground surface in a radiation shield) were also measured. Sensors for ground surface temperature measurements were generally placed underneath a plant or about 1 cm below the surface for bare ground to avoid solar heating. The sensors for air and ground surface temperature measurements were also painted white to minimize solar heating. Both air and soil temperatures were recorded to the nearest 0.01°C with an estimated accuracy of ±0.1°C. Air temperature data were also obtained from the standard weather service stations. Daily mean air temperatures at the national Weather Service stations were calculated as the average of daily maximum and minimum air temperatures.

[17] For depths greater than 1.0 m, temperatures were logged by using a high-precision thermistor and cable. The resistances of the thermistor and cable were measured by a Digital MultiMeter and were converted into temperatures in the laboratory. The sensitivity of the measurements was ±1 ohm. The rate of temperature change with resistance was about 3×10^{-3}°C ohm^{-1}, i.e., the sensitivity of thermistor was about $\pm 3 \times 10^{-3}$°C.

3.2. Timing and Duration

[18] Seasonal snow cover fluctuates widely, both spatially and temporally. Variations in the timing and duration of the seasonal snow cover result in variations in the ground thermal regime, both in magnitude and in its vector (cooling or warming). The timing of the seasonal snow cover refers to the date of snow cover establishment in late autumn/early winter and the date of snow cover disappearance in spring, whereas the duration of the seasonal snow cover refers to the time period when snow covers the ground surface. Recent studies indicate that the duration of seasonal snow cover over north central and northwest Asia has increased by about 4 days per decade based on historical records from 1937 through 1994, because of an earlier first snow date in autumn and also a later last snow data in spring [*Ye*, 2001]. The date of snow disappearance in spring on the Alaskan North Slope varies significantly from year to year, up to a month [*Kane et al.*, 1991; *Zhang et al.*, 1997]. The spring snowmelt in northern Alaska has advanced by about 8 days on average since the mid-1960s [*Stone et al.*, 2002]. The advance appears to be a consequence of decreased snowfall in winter, followed by warmer spring conditions.

[19] Snow cover has a cooling effect on the ground surface when snow is relatively thin and air temperature fluctuates around 0°C at the beginning of the snow accumulation in autumn. This cooling effect occurs because fresh snow has a relatively high surface albedo when the solar elevation is still relatively high in autumn. The duration of the cooling effect may be very short, though, and thus it may have very little effect on the mean annual ground surface temperature. As the air temperature decreases and snow thickness increases with time, the insulating effect of snow cover becomes a dominant factor, preventing the ground from cooling. The duration of snow insulating effect may vary from a few weeks to several months, depending upon the geographical locations. During spring when snow melts, the ground surface temperature is a few degrees lower than the air temperature, mainly because a considerable fraction of the solar radiation is reflected and consumed during snow melting. Snowmelt is generally a very rapid process, usually ranging from a few days to a couple of weeks in the Arctic and sub-Arctic [*Weller and Holmgren*, 1974; *Hinzman et al.*, 1991; *Kane et al.*, 1991; *Stone et al.*, 1996; *Zhang et al.*, 1996a; *Stone et al.*, 2002]. This cooling effect of snowmelt on the ground surface can be significant in the short term but relatively small in terms of its effect on the mean annual ground surface temperature.

[20] The effect of the seasonal snow cover on the ground thermal regime may be different depending on the time period of the interest. On a daily basis, snow cover results in either warmer or cooler ground surface depending upon the variations in air temperature and the prior thermal history of the ground surface [*Zhang et al.*, 1997]. Figure 2 shows the difference between daily average ground surface and air temperatures at Franklin Bluff, northern Alaska, for the period from the middle of March 1987 through the middle of March 1988. Throughout this study, when we say the difference (ΔT) between ground surface temperature (T_g) and air temperature (T_{air}), we mean $\Delta T = T_g - T_{\mathrm{air}}$. Generally, the difference varied from about −1°C to 5°C during the summer months (June through September), while during the cold season when the ground surface is covered by snow, the difference could range from about −10°C to 20°C. The ground surface temperature was often about a few degrees higher than the air temperature during the surface snow-covered period, while the largest daily value recorded was 36.3°C higher in the winter of 1988 at Franklin Bluffs on the North Slope of Alaska. When a warm air mass was advected into the region, the snow cover would prevent the ground surface from warming (negative peaks from December through March in Figure 2). The ground surface temperature was occasionally a few degrees lower than the air

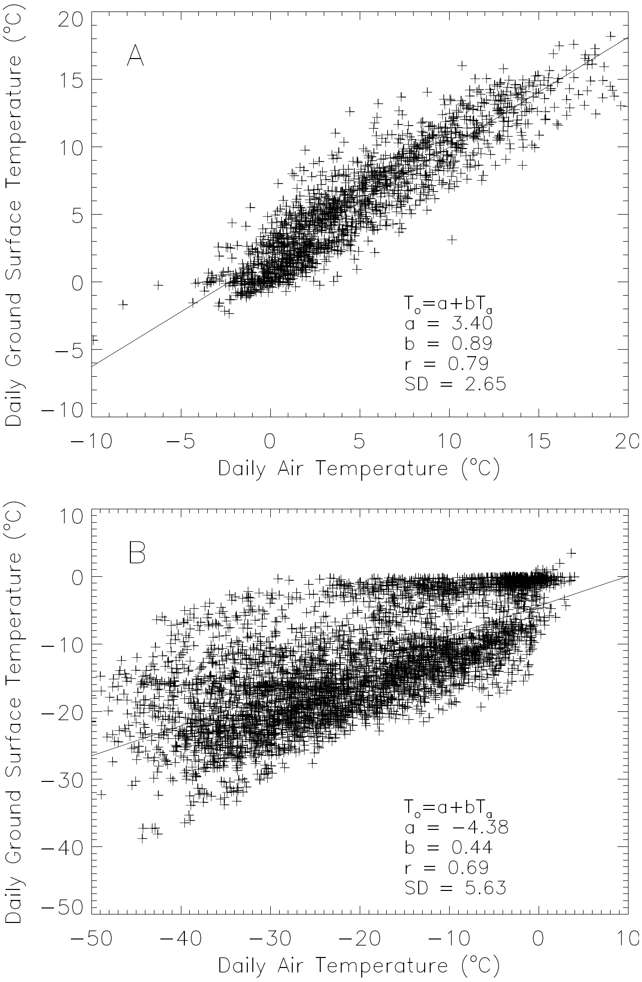

Figure 3. Relationships between the daily mean air and ground surface temperatures for (a) snow-free and (b) snow-covered land surfaces at West Dock, Deadhorse, and Franklin Bluffs, North Slope of Alaska, for the period from 1987 through 1992 [*Zhang et al.*, 1997]. T_o stands for daily mean ground surface temperature in °C, T_a stands for daily mean air temperature, *a* and *b* are correlation constants, *r* is the correlation coefficient, and SD is the standard deviation of the measured temperature from the predicted values using the linear equations. Copyright John Wiley and Sons Ltd. Reproduced with permission.

temperature, up to −13.8°C as recorded in the winter of 1992 at West Dock, Prudhoe Bay. During late April and early May 1987 (Figure 2) the ground surface temperature was consistently a few degrees lower than the air temperature. This cooling effect is mainly due to the snowmelt impact.

[21] Figure 3 illustrates the daily average air temperature versus ground surface temperature during the surface snow-free (Figure 3a) and snow-covered (Figure 3b) periods from 1987 through 1992 in northern Alaska [*Zhang et al.*, 1997]. During the snow-free period, there is a good linear correlation between air and ground surface temperatures, with a correlation coefficient of 0.89 and a standard deviation of 2.65°C. During the snow-covered period this linear correlation weakens, with a correlation coefficient of 0.69 and a standard deviation of 5.63. For example, for a daily average ground surface temperature of −15°C the daily average air temperature ranges from −50°C to about 0°C (Figure 3b). On the other hand, for a daily mean air temperature of −15°C the daily mean ground surface temperature could vary from −25°C to 0°C. This demonstrates that on a daily basis, snow can have either positive or negative influence on the ground surface temperature through its insulating effect.

[22] Snow cover can also reduce the amplitude of the daily ground surface temperature (Figure 4). During the snow-free period the amplitude of the daily ground surface temperature had the same amplitude of daily air temperature, ranging from 4°C to 8°C. During the snow-covered period the amplitude of daily ground surface temperature was reduced to lower than 2°C, most of the time even less than 1°C (Figure 4). The sharp change in the amplitude of the daily ground surface temperature can be used to detect the first and last day of snow on the ground (Figure 5), and the results are consistent with results from the National Weather Service records [*Zhang et al.*, 1997].

[23] On a monthly basis, snow cover can have either positive or negative impact on the ground surface temperature depending upon the time of the year. Figure 6 shows the difference between the mean monthly ground surface and air temperatures over the period from 1987 through 1992 at three sites in northern Alaska [*Zhang et al.*, 1997]. The insulation effect was greater during the early winter, reaching its maximum in November at all three sites, because of the low density, and thus low thermal conductivity, of fresh snow. Latent heat released during the active layer freeze-up process also helps to keep the ground surface from cooling. The mean ground surface temperature in November could be up to 20°C warmer than the mean monthly air temperature at Franklin Bluffs on the North Slope of Alaska. The ground surface temperature was a few degrees lower than air temperature in May (Figure 6) because of the effect of snowmelt. *Sokratov and Barry* [2002] further demonstrated that the effect of snow depth on the soil surface energy balance could be divided into four

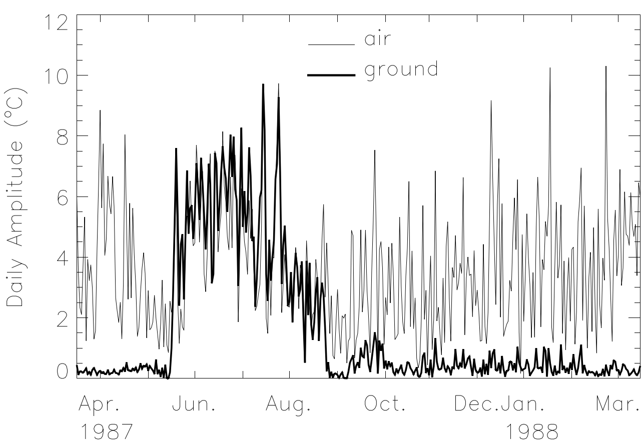

Figure 4. Daily amplitude of air temperature (thin line) and ground surface temperature (thick line) at Framklin Bluffs, North Slope of Alaska, for the period from mid-March 1987 through mid-March 1988 [*Zhang et al.*, 1997]. Copyright John Wiley and Sons Ltd. Reproduced with permission.

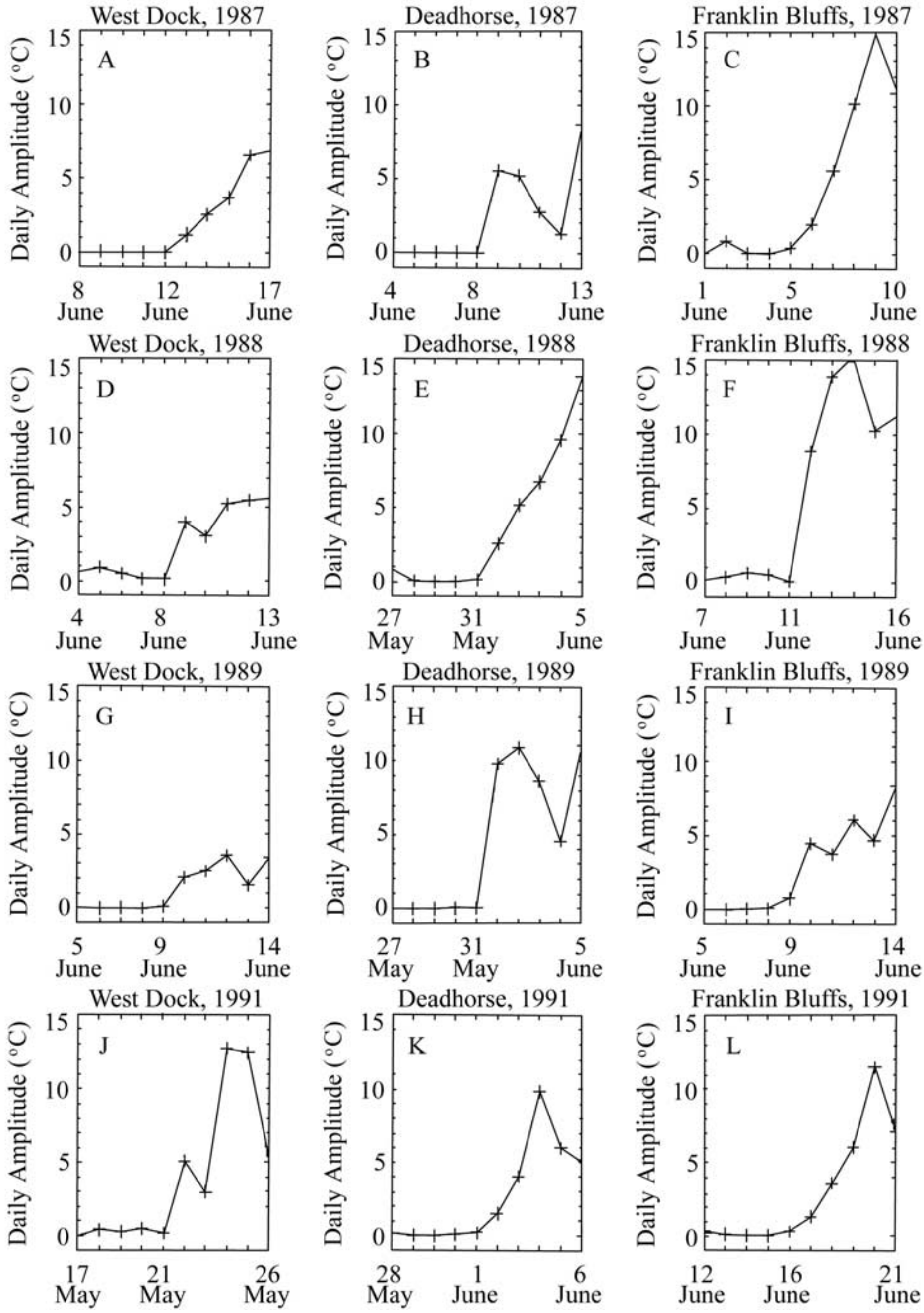

Figure 5. Abrupt increase in the daily amplitude of the ground surface temperature due to the disappearance of the seasonal snow cover at West Dock, Deadhorse, and Franklin Bluffs on the North Slope of Alaska. The sharp change in the daily amplitude of the ground surface temperature can be used to determine the first and last date of snow on the ground, and results are consistent with other available data [*Zhang et al.*, 1997]. Copyright John Wiley and Sons Ltd. Reproduced with permission.

major intraseasonal stages. On the basis of data and information from Barrow, Alaska, they concluded that snow depth has a direct effect on the soil surface energy balance from the date of the active layer freeze-up to the date when snow depth reaches maximum (stages II and III of *Sokratov and Barry* [2002]).

[24] On an annual basis, seasonal snow cover has a positive impact on the ground surface temperature in the Arctic. For example, for a mean annual air temperature of about −13°C on the North Slope of Alaska, mean annual ground surface temperature ranged from −5°C to −10°C (Figure 7) due mainly to the effect of the seasonal snow cover. Assuming the temperature difference between the mean annual ground surface temperature and the mean annual air temperature is 1.1°C, which is the average value of the temperature difference during the snow-free period in summer in northern Alaska, seasonal snow cover can increase the mean annual ground surface temperature by 4°C to 9°C in this region (Figure 7) [also see *Zhang et al.*, 1997].

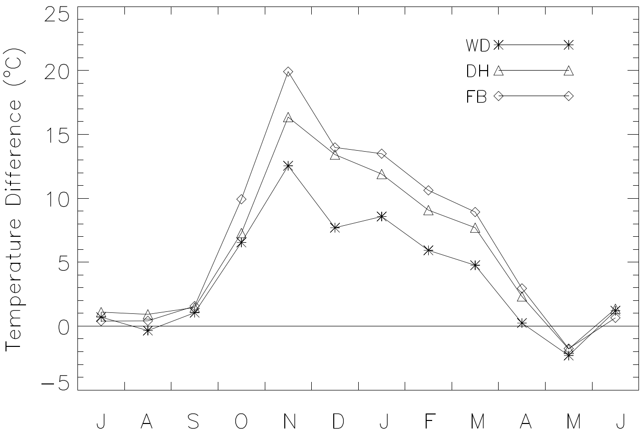

Figure 6. Temperature difference between the mean monthly ground surface and air temperatures at West Dock (WD), Deadhorse (DH), and Franklin Bluffs (FB), North Slope of Alaska, for the period from 1987 through 1992 [*Zhang et al.*, 1997]. The solid horizontal line is 0°C reference. Copyright John Wiley and Sons Ltd. Reproduced with permission.

[25] Results from numerical modeling indicate that mean annual ground surface temperatures are strongly influenced by an accumulation of snow in autumn and early winter [*Goodrich*, 1982]. Further sensitivity analyses demonstrate that for the same air temperature and snow conditions, early snow cover in autumn with early snowmelt in spring provides the most favorable condition for snow cover to influence the ground thermal regime, increasing the mean annual ground surface temperature by up to 6°C. The combination of late snow cover in autumn/winter with late snowmelt in spring is the least favorable condition, leading to a decrease in mean annual ground surface temperature by up to 3°C, depending upon the specific conditions. The effect of early snow cover with late snowmelt and late snow cover with early snowmelt on the ground thermal regime would change the mean annual ground surface temperature by 2.0°C and −1.5°C, respectively, compared with the average conditions at Barrow, Alaska [*Zhang*, 1993; *Ling and Zhang*, 2003]. Changes in spring snowmelt dates alone have relatively little impact on the mean annual ground surface temperature [*Goodrich*, 1982; *Zhang*, 1993]. *Ling and Zhang* [2003] found that delaying the first snow date in autumn and the last snow date in spring by 10 days at Barrow, Alaska, results in a decrease of the maximum ground surface temperature by up to 9.0°C and a decrease of the mean annual ground surface temperature by 0.7°C. Advancing the last snow date in spring by 10 days leads to an increase of the ground surface temperature by 6.6°C and an increase of the mean annual ground surface temperature by 0.2°C. Variations in the timing and duration of seasonal snow cover also have an influence on active layer thickness, but the effect is very limited.

3.3. Snow Depth

[26] Snow depth is one of the major factors that control the magnitude of the influence of the seasonal snow cover on the ground thermal regime in cold regions. In the Russian literature it was observed [*Kudryavtsev*, 1992] that when snow cover is relatively thin with high albedo, snow cover results in a cooler soil surface. As snow thickness increases, the insulating effect of snow cover increases and results in a warmer soil surface. The insulating effect reaches maximum when snow is at its optimal thickness (about 40 cm). Then the insulating effect (warming) will decrease as snow thickness continues to increase. If the snow cover is thick enough so that snow can last until late spring or summer, snow cover may overall have a cooling effect because of its albedo and latent heat effect. Further increase in snow thickness may form permanent snow patches for many years or glaciers. In this case, seasonal air temperature variation may not penetrate through snow patches or glaciers, and snow may have a warming impact again.

[27] It has been reported in the Russian literature that, on average, an increase in snow thickness by 5 to 15 cm leads to a 1°C increase in mean annual ground temperature [*Yershov*, 1998]. With a sufficient thickness of the seasonal snow cover the mean annual ground surface temperature can be positive when the mean annual air temperature is low, up to −6°C to −8°C. For example, the mean annual air

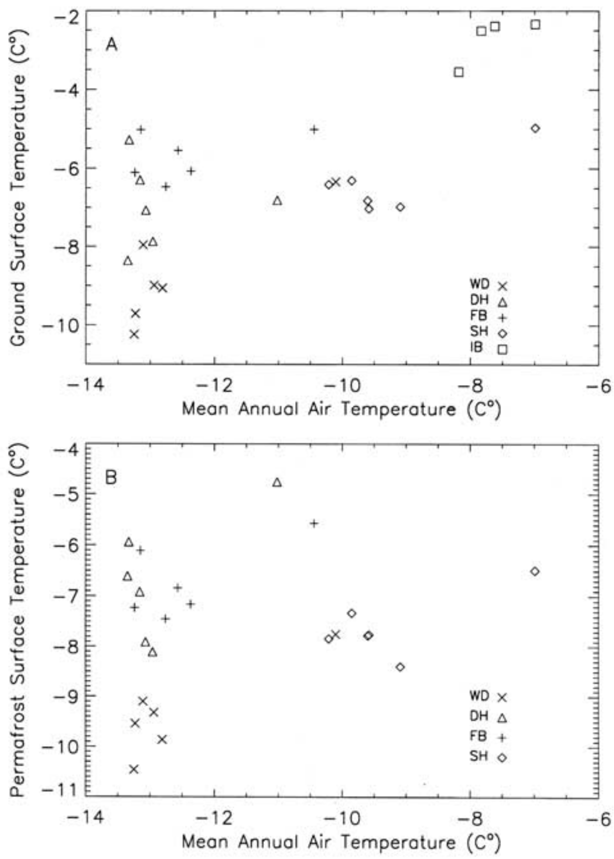

Figure 7. Relationships (a) between mean annual air and mean annual ground surface temperatures and (b) between mean annual air and near-permafrost surface temperatures for the period from 1987 through 1992 at West Dock (WD), Deadhorse (DH), Franklin Bluffs (FB), Sagwon (SH), and Toolik Lake (IB) on the North Slope of Alaska. From *Zhang et al.* [1997].

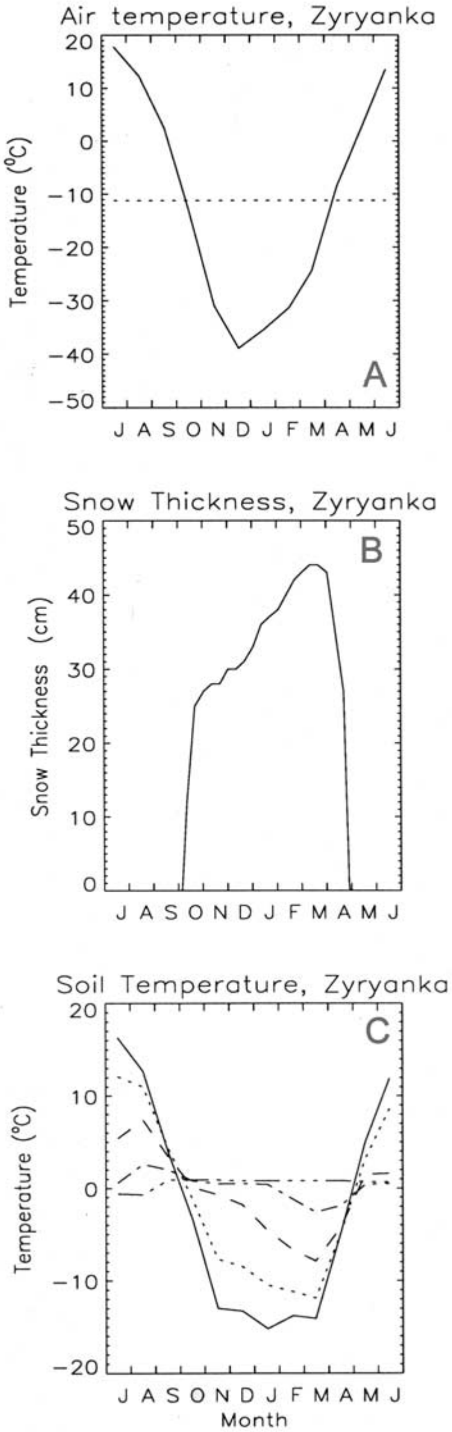

Figure 8. Relationship between (a) air temperature, (b) snow cover thickness, and (c) soil temperatures at various depths at Zyryanka (65°N, 150°E, 43 m above sea level), Russia. In Figure 8a the solid line represents mean monthly air temperature, and the dashed line represents the mean annual air temperature.

temperature at Vorkuta, Russia, was about −6.4°C, while the mean annual ground surface temperature was 1.0°C [*Pavlov*, 1973]. Figure 8 shows seasonal variations of air temperature (Figure 8a), snow cover thickness (Figure 8b), and soil temperatures at various depths (Figure 8c) at Zyryanka, Russia. It demonstrates that the mean annual air temperature there is about −11°C, with an amplitude of about 29°C, and snow covers the surface for more than 6 months (from the beginning of October to the end of April) with maximum snow thickness of about 45 cm. The mean annual ground temperature at a depth of 3.2 m is only a few tenths of a degree below 0°C. *Zhang et al.* [1997] reported that the mean annual air temperature at Prudhoe Bay, Alaska, is about −12.4°C with an amplitude of about 16°C, and where snow covers the surface from late September or early October to late May or early June with average maximum snow thickness of about 20 cm, the permafrost surface temperature is about −9°C with permafrost thickness of greater than 600 m. Numerical modeling results indicate that the rate of the mean annual ground surface temperature increases with the increase of the maximum snow depth to about 0.1°C cm^{-1} at the maximum snow depth of 15 cm. This rate decreases as the maximum snow depth increases [*Zhang*, 1993].

[28] *Sturm and Holmgren* [1994] demonstrate that within a horizontal distance of about 1 to 1.5 m, snow depth varied from about 35 to 10 cm because of the local microrelief impact, and the snow-ground interface temperature changed from about −5°C to −15°C, respectively. M. Sturm and J. Holmgren (unpublished research, 2001) conducted detailed field measurements of the snow-ground interface temperature and snow depth at Ivotuk, Alaska, on 17 November 1998. This research clearly shows that the snow-soil interface temperatures were much lower on the top of hills where snow was relatively thin than those on the slope where snow was relatively thick (Figure 9). The snow-soil interface temperature increases as snow depth increases, and snow depth effect becomes less pronounced when it is greater than 40 cm (Figure 10). Measurements in the Mackenzie Delta show similar results [*Smith*, 1975]. Field measurements also indicate that there is a characteristically steep temperature gradient immediately below the snow surface, with a marked inflection at a depth of 50 to 60 cm, below which temperatures are fairly uniform [*Nicholson and Granberg*, 1973]. These results imply that variations in snow depth are more critical for ground temperatures where the seasonal snow cover is relatively thin.

[29] In the northeastern region of China, field measurements indicate that mean ground surface temperatures were up to 9°C higher with snow cover thickness on the ground ranging from 10 to 25 cm than on the bare surface (snow was removed after snowfall) for the period from November 1977 through March 1978 [*Zhou et al.*, 2000]. *Liang and Zhou* [1993] conducted a detailed field investigation on snow distribution, thickness, and surface and subsurface soil temperatures in the Amuer region during the winter of 1991–1992. Their results indicate that the seasonal snow cover with thickness ranging from 21 to 36 cm resulted in mean annual ground surface temperatures about 2.8° to 5.0°C warmer than that over the snow-free ground surface in the study area.

[30] *Goodrich* [1982] conducted a detailed numerical simulation on the general features of the impact of the

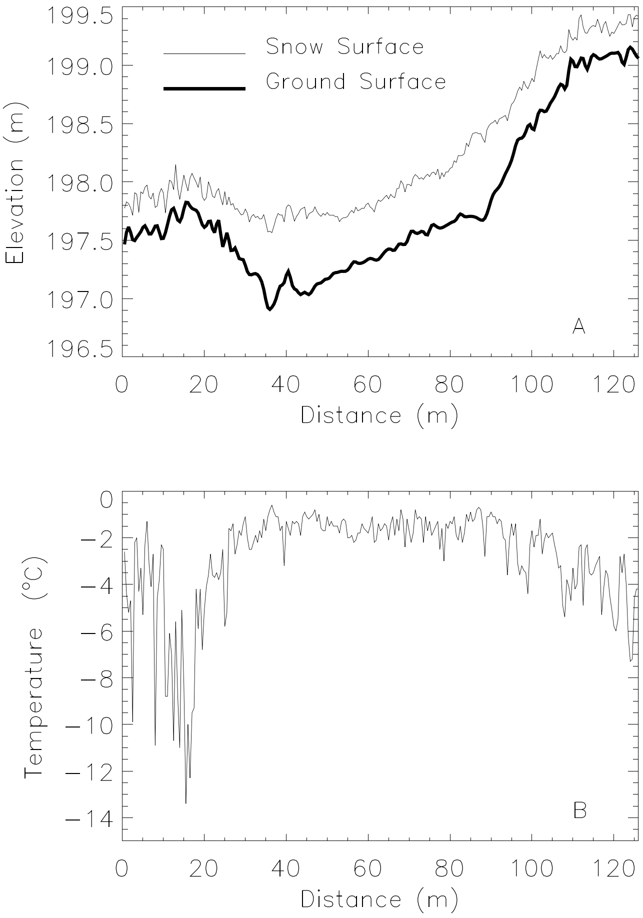

Figure 9. Variations of (a) snow thickness and (b) snow-ground interface temperatures with local microrelief at Ivotuk, Alaska, taken on 17 November 1998. Data were provided by M. Sturm and J. Holmgren at Cold Region Research and Engineering Laboratory (CRREL), Fairbanks, Alaska.

seasonal snow cover on long-term, periodic, steady state equilibrium ground temperatures using a heat conduction model with phase change. For the mean annual surface temperature of −10°C with an amplitude of 20°C (snow surface when snow is present and bare soil surface when snow is absent), a doubling of snow depth from 25 to 50 cm resulted in a minimum ground surface temperature rise by about 7°C (from −19.7°C to −12.3°C) and the mean annual ground surface temperature rise by 3.6°C (from −7.0°C to −3.4°C). If there was 50 cm of snow built up within 30 days in autumn, the minimum ground surface temperature would be only −2.8°C, the mean annual ground surface temperature would be +1.1°C, and permafrost would degrade. Using the same numerical model, sensitivity analysis indicates that for the same air temperature condition and snow cover accumulation process, mean annual ground surface temperature increases as the maximum snow depth increases [*Zhang*, 1993]. The active layer freeze-up date would also be delayed as the maximum snow depth increased. For example, the active layer freeze-up date would be delayed about 3 weeks for the maximum snow depth varying from 15 to 50 cm.

3.4. Snow Density and Structure

[31] The insulating effects of the seasonal snow cover on the ground thermal regime result from its low thermal conductivity, which, in turn, depends on snow density and snow cover structure. Snow density ranges from less than 100 kg m^{-3}, for fresh snow and depth hoar, to greater than 600 kg m^{-3}, for melting snow and wind slab. Such a spectrum of snow density variation has a significant impact on snow thermal properties (Figure 11) and thus its insulating effect on the ground thermal regime.

[32] Using the average snow and climatic conditions along the Alaskan Arctic, a sensitivity study indicates that changes in the thermal conductivity of snow cover have a significant impact on the ground surface, active layer, and permafrost temperatures. Along the Alaskan Arctic coast (within about 20 km from the coast) the mean annual air temperature is −12.4°C with an amplitude of about 16.0°C [*Zhang et al.*, 1997]. Snow covers the ground surface for about 9 months, from the middle of September to the beginning of June. Snow depth reaches its maximum of about 35 cm in late April. *Goodrich*'s [1982] model was used to investigate the impact of changes in the thermal conductivity of snow cover on the ground thermal regime. The results demonstrate that changes in the thermal conductivity of snow cover from 0.7 W m^{-1} K^{-1} to 0.1 W m^{-1} K^{-1} would lead to an increase in the mean annual ground surface temperature of about 5.5°C from −6.5°C to −1.0°C. The minimum ground surface temperature would increase from −16.3°C to −3.7°C (Figure 12). Changes in the thermal conductivity would also reduce the annual amplitude of the ground surface temperature from 10°C to 3.6°C and delay the active layer freeze-up date by up to 4 months. For the thermal

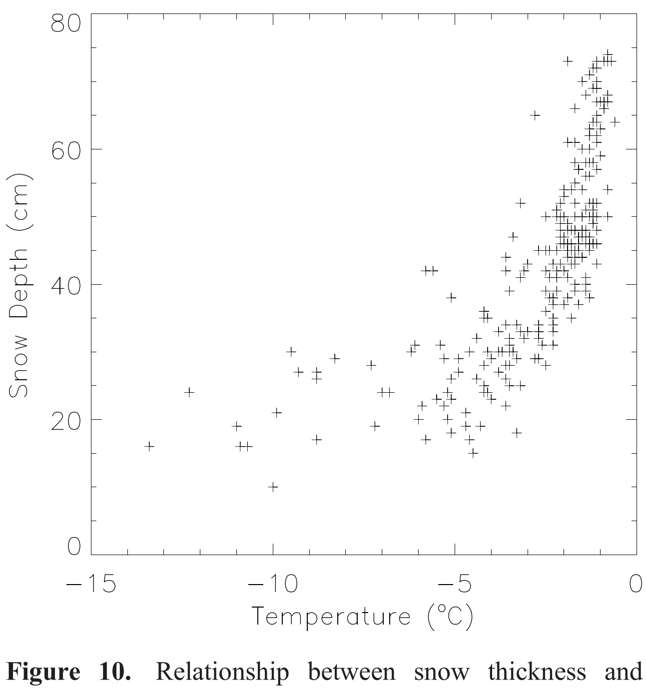

Figure 10. Relationship between snow thickness and snow-ground surface interface temperature at Ivotuk, Alaska, taken on 17 November 1998. Data were provided by M. Sturm and J. Holmgren at CRREL, Fairbanks, Alaska.

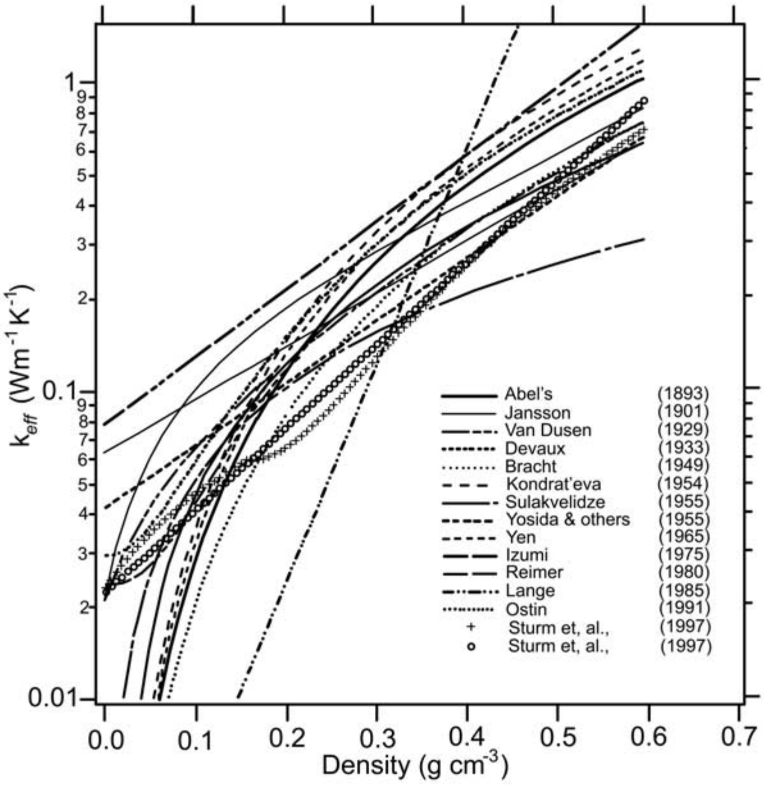

Figure 11. Published regression equations of thermal conductivity versus density [*Sturm et al.*, 1997]. Reprinted with permission of International Society of Glaciology.

conductivity of 0.7 W m^{-1} K^{-1} the freeze-up date of the active layer is around the first week of December, while for the thermal conductivity of 0.1 W m^{-1} K^{-1} the freeze-up date is delayed to the middle of April. Results from *Goodrich*'s [1982] study indicate that for a snowpack thickness of about 50 cm with an effective thermal conductivity of about 0.2 W m^{-1} K^{-1} the freeze-up date of the active layer could be delayed to late March.

[33] Field studies show that the seasonal snow cover in the Arctic and sub-Arctic consists of layers with distinct physical and thermal properties. *Benson and Sturm* [1993] reported that the tundra snow consists of hard, high-density, wind-packed layers at the top with coarse and low-density depth hoar layers at the base. The density of the hard wind-packed layers (wind slab) varies from 0.4 to 0.5 g cm^{-3} with the grain size ranging from 0.5 to 1.0 mm; density for the depth hoar layers varies from 0.1 to 0.25 g cm^{-3} with the grain size ranging from 5.0 to 10.0 mm. On the basis of Figure 11 the thermal conductivity can vary from less than 0.01 W m^{-1} K^{-1} to greater than 1.0 W m^{-1} K^{-1}. The depth hoar fraction can be over 50% of this entire snowpack [*Sturm and Johnson*, 1991; *Sturm and Benson*, 1997]. *Sturm and Johnson* [1991] also reported that the depth hoar fraction for the Fairbanks snow cover (taiga snow) can be up to 80%. The thermal conductivity of the depth hoar varies from 0.026 to 0.105 W m^{-1} K^{-1}, with an average value of 0.063 W m^{-1} K^{-1} [*Sturm and Johnson*, 1992; *Sturm et al.*, 1997]. This is about one fifth to one twentieth the value of the high-density wind slab (Figure 11).

[34] *Zhang et al.* [1996c] investigated the effect of changes in the depth hoar fraction in the seasonal snow cover on the ground thermal regime in cold regions by modifying *Goodrich*'s [1982] original model. The snow cover was treated as a single layer in the numerical model simulation. However, the effective thermal properties of this single layer were calculated separately using a serial model that includes the wind slab and depth hoar layer of the

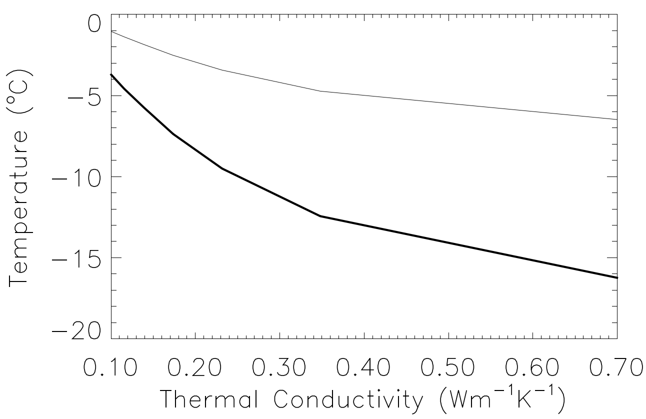

Figure 12. Impact of the effective thermal conductivity of the seasonal snow cover on ground surface temperatures. The thin curve represents changes of mean annual ground surface temperature with the effective thermal conductivity of snow, while the thick curve represents changes in minimum ground surface temperatures. From *Zhang et al.* [1996c].

seasonal snow cover. The results show that the effective thermal conductivity and effective thermal diffusivity of the snowpack decrease nonlinearly with increases in the depth hoar fraction, although the effective volumetric heat capacity decreases linearly with an increase of the depth hoar fraction. The effect of the depth hoar thermal conductivity on the effective thermal conductivity and diffusivity of the entire snowpack is much greater than the effect of the wind slab thermal conductivity, especially when the depth hoar fraction is relatively high.

[35] Sensitivity analyses show that for the average snow and climate conditions along the Alaskan Arctic coast as described above, changes in the depth hoar fraction from 0.0 (no depth hoar) to 0.6 would result in an increase of the ground surface temperature by 12.8°C, an increase of the mean annual ground surface temperature by 5.5°C, and a decrease of the annual amplitude of the ground surface temperature by 6.3°C (Figure 13) [*Zhang et al.*, 1996c]. Sensitivity analyses were also conducted for the seasonal frost case with the average snow and climatic conditions as follows: a mean annual air temperature of 2.0°C and an annual amplitude of 25.0°C, with snow cover from the middle of November to the end of April, and a maximum snow thickness of 35 cm in the middle of March. Changes in the depth hoar fraction from 0.0 (no depth hoar) to 0.6 would result in an increase of the minimum ground surface temperature by 8.4°C, an increase of the mean annual ground surface temperature by 2.4°C, and a decrease in the annual amplitude of the ground surface temperature by 4.3°C. For the permafrost case, changes in the depth hoar fraction from 0.0 to 0.6 delay the freeze-up date of the active layer up to 4 months, while for the seasonal frost case the freezing depth decreases from 1.15 to 0.18 m, or more than 80%. Certainly, these are likely extreme ranges.

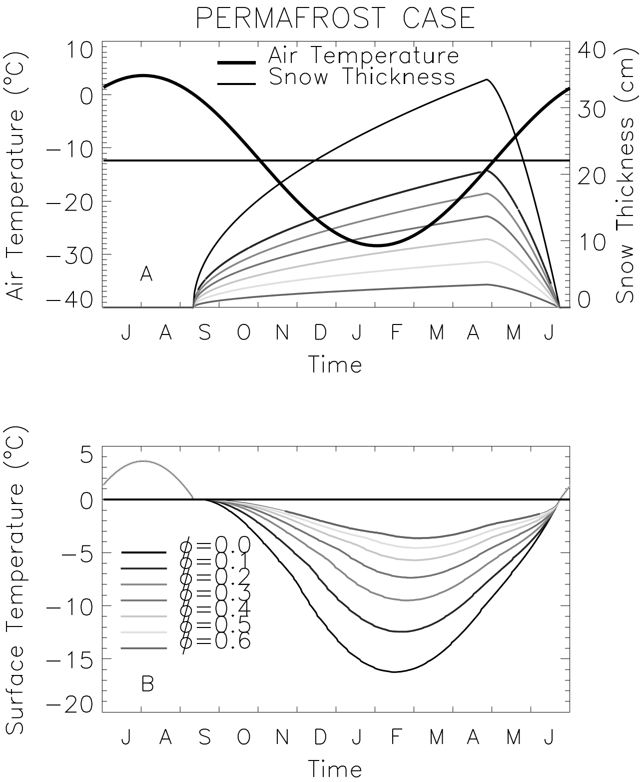

Figure 13. Impact of changes in the depth hoar fraction of the seasonal snow cover on ground surface temperature. (a) Input air temperature, mean annual air temperature (solid horizontal line), and snow and depth hoar thickness for the case along the Alaskan Arctic coast. (b) Output ground surface temperatures with variations of the depth hoar fraction in the seasonal snow cover. Line types in Figures 13a and 13b are consistent. From *Zhang et al.* [1996c]. See color version of this figure at back of this issue.

4. IMPACT OF THE SEASONAL SNOW COVER ON SEASONAL FREEZE/THAW AND PERMAFROST DEVELOPMENT

[36] As shown in section 3, snow cover has a significant impact on ground temperatures. In turn, changes in ground temperatures have a great impact on the plant species and ecosystem processes of numerous biomes. *Walker et al.* [1997] provide extensive information regarding tundra snow-vegetation interactions and mainly focus on the effect of snow on vegetation toward a unified hierarchical understanding of species-, community-, landscape-, and biome-level responses to varying snow regimes. On the other hand, seasonal snow cover combined with surface microrelief and vegetation has a profound impact on distribution, temperature, and thickness of seasonal freeze/thaw and permafrost in cold regions.

[37] Frozen ground can be divided into two major categories: seasonally frozen ground, where ground freezes annually, and perennially frozen ground or permafrost, where ground (soil or rock and included ice and organic material) remains at or below 0°C for at least 2 years [*Permafrost Subcommittee*, 1988]. Seasonal freezing refers to the process of freezing from the top of unfrozen ground, and the layer of seasonal freezing is underlain by unfrozen ground. Seasonal thawing refers to thawing from the top of frozen ground, and the layer of seasonal thawing is always underlain by permafrost.

[38] Geographically, the duration and thickness of the seasonal frozen ground increase from south to north in the Northern Hemisphere and from low to high elevation, increasing into permafrost regions. On the basis of its area extent the permafrost region can be divided into four categories: continuous, discontinuous, sporadic, and isolated permafrost zones. The definition of these permafrost zones is descriptive, and the boundary between any adjacent two permafrost zones is generally ambiguous. According to recently published information [*Brown et al.*, 1997; *International Permafrost Association Data and Information Working Group*, 1998; *Zhang et al.*, 1999a] (Figure 14) the continuous permafrost zone is defined as a region where permafrost underlies more than 90% of the total area; the discontinuous permafrost zone is defined as a region where permafrost underlies 50% to 90% of the total area; the sporadic permafrost zone is defined as a region where permafrost underlies 10% to 50% of the total area; and the isolated permafrost zone is defined as a region where

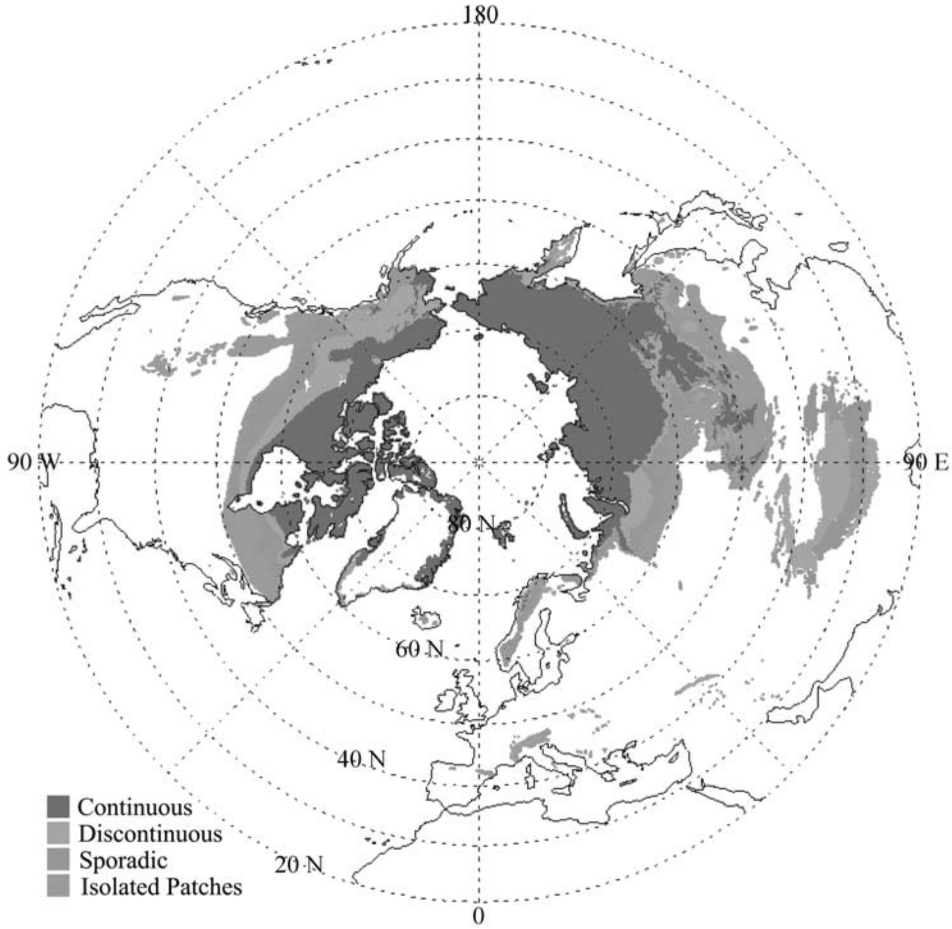

Figure 14. Permafrost distribution in the Northern Hemisphere. Data were obtained by *Brown et al.* [1997] and simplified by *Zhang et al.* [1999a]. Reprinted with permission of V. H. Winston and Son, Inc. See color version of this figure at back of this issue.

less than 10% of the area is underlain by permafrost. About 57% of the exposed land surface freezes and thaws seasonally [*Zhang et al.*, 2003], while permafrost zones occupy about 23.9% of the exposed land surface in the Northern Hemisphere [*Brown and Haggerty*, 1998; *Zhang et al.*, 1999a, 2000].

4.1. Continuous Permafrost

[39] Seasonal snow cover, combined with other environmental factors, has a significant influence on permafrost temperatures and thickness in the continuous permafrost zone. In northern Alaska, within about 120 km from the Arctic coast, the mean annual air temperature is nearly constant at about −12.4°C, while the mean annual permafrost surface temperature varies from −9.1°C at West Dock to about −5.0°C at Happy Valley, a 4.1°C increase, and permafrost thickness decreases from over 600 m along the coast to about 350 m in the interior of the North Slope of Alaska [*Zhang et al.*, 1997] (Figure 15). Except for the impact of the length of thaw season and summer air temperature, an effect which is relatively smaller in magnitude, the interactions of wind, microrelief, and vegetation with the seasonal snow cover and variations of physical (such as density and structure) and thermal properties of the seasonal snow cover are the major factors affecting permafrost temperatures during the 8- to 9-month winter on the North Slope of Alaska [*Zhang et al.*, 1997]. Variations of wind, microrelief, and vegetation from the Arctic coast to the foothills of the Brooks Range are substantial. The Arctic coast is characterized by strong winds, relatively flat surface with low central polygons, and poorly developed vegetation. These conditions are favorable for snow to be blown away or very hard packed. Either of these cases would reduce the snow insulating effect that results in lower permafrost temperatures along the Arctic coast. In contrast, the Arctic inland and Arctic foothills are characterized by lower wind speeds, very rough surface with tussocks, troughs, and depressions and well-developed vegetation with the willow communities reaching heights of about 0.5 to 1.1 m. Snow can be interrupted and trapped by vegetation and rough surface, increasing the insulating effect and permafrost temperatures. Recent study indicates that since 1950, area coverage of tall shrubs within Alaska's North Slope tundra has increased 1.2% per decade (from 14 to 20% area coverage) [*Sturm et al.*, 2001]. Because of the fact that more snow would be accumulated beneath shrubs, the increased snow insulation effect might account for the permafrost warming in northern Alaska.

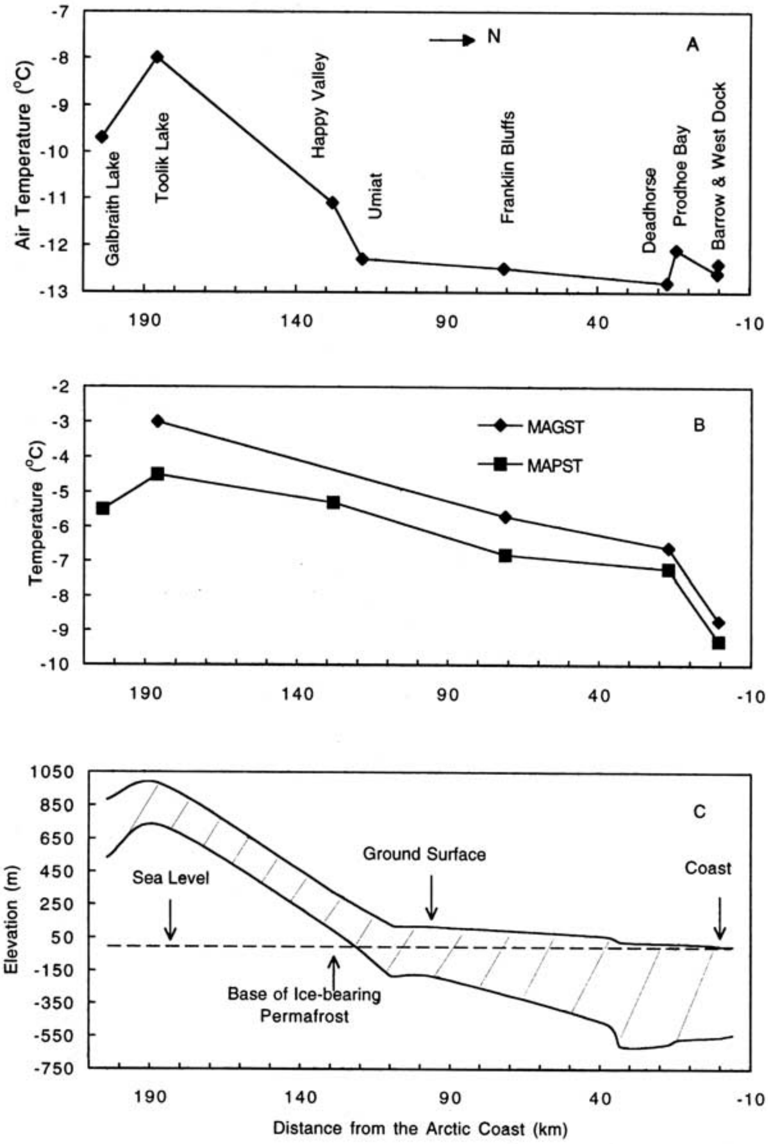

Figure 15. Variations of (a) mean annual air temperature, (b) ground surface and permafrost surface temperatures, and (c) permafrost thickness with distance from the Arctic coast on the North Slope of Alaska. MAGST is mean annual ground surface temperature; MAPST is mean annual permafrost surface temperature [*Zhang et al.*, 1997]. Copyright John Wiley and Sons Ltd. Reproduced with permission.

[40] In the Canadian Arctic Archipelago the effect of snow cover is considerably diminished from its influence farther south in the permafrost region because of the lower snowfall, higher density, and very uneven distribution because of drifting [*Brown*, 1972; *Brown and Pewe*, 1973]. *Menard et al.* [1998] reported similar results for the combined impact of vegetation and seasonal snow cover on ground surface temperatures near Umiujaq, eastern Hudson Bay, Canada. Their results indicate that ground surface temperature was lower where both thin vegetation (moss and lichen covers) and snow cover exist, while ground surface temperature was higher where ground is covered by shrubs and trees with thicker snow cover.

[41] *Smith et al.* [1998] reported that permafrost temperatures in the boreal forest and taiga ecozones show similar values (varying from −1°C to −2°C), while in the tundra ecozone, permafrost temperatures decline markedly in northwestern Canada (about −6°C). They attributed this dramatic change in permafrost temperatures to the lack of deep snow cover over upland soil surface and a longer cold season in the tundra environment. *Nixon and Taylor* [1998] reported that a larger air temperature thawing index is required for the boreal forest than for tundra to achieve a similar active layer thickness because of the combined impact of surface vegetation and generally deeper snow cover in the Mackenzie Valley, compared to the Mackenzie Delta, in northwestern Canada.

4.2. Discontinuous Permafrost

[42] Seasonal snow cover has a significant impact not only on permafrost temperatures and thickness but also on permafrost distribution in the discontinuous permafrost

zone. A comprehensive and systematic investigation of the impact of the seasonal snow cover on permafrost temperature, thickness, and distribution in the discontinuous permafrost zone has been conducted near Schefferville in northern Quebec since the late 1950s. Permafrost was reported to be more than 100 m in thickness [*Annersten*, 1966]. Further studies indicated that snow cover is a dominant factor in controlling permafrost distribution at this area [*Gold and Lachenbruch*, 1973]. *Granberg* [1973] reported that the distribution of discontinuous permafrost near Schefferville (54°48′N, 66°49′W) is closely related to the patterns of seasonal snow cover accumulation. Deep snow cover prevents the development of permafrost even if the mean annual air temperature is about −4.5°C in the region.

[43] Some studies tried to establish a simple relationship between mean snow depth and ground temperatures at various depths. In general, no permafrost was found in areas where the thickness of snow cover exceeded about 40 cm, although it is doubtful that strict application of the less than 40 cm snow depth as an indicator of permafrost is applicable in all cases [*Brown and Pewe*, 1973]. After extensive field measurements, *Nicholson and Granberg* [1973] concluded that the depth of snow cover corresponding to a mean annual temperature of 0°C at various depths (varying from 1.5 to 45 m) ranges from 70 to 75 cm. This implies that permafrost might develop where the average snow depth for the area of influence is less than 70 to 75 cm. On the basis of extensive field measurements a simple model of the linear relationship between average snow depth and soil temperatures at certain depths was established and used to predict the distribution of permafrost in the region [*Nicholson and Granberg*, 1973]. Later, a few modified simple models were developed with some improvement [*Jones*, 1976; *Nicholson*, 1978a, 1978b].

[44] Recent studies illustrate that there is great complexity in heat transfer through forest and tundra snow cover at Schefferville, Quebec [*Desrochers and Granberg*, 1988; *Granberg*, 1988]. A wide range of factors complicates the simple relationship between snow depth and ground temperature that has been used in permafrost prediction at Schefferville. However, these findings do not necessarily render the simple relationship invalid, although they do open a new perspective on the role of seasonal snow cover in controlling the ground temperature and the spatial distribution of permafrost at Schefferville and elsewhere. For example, a critical depth of about 50 cm snow cover or greater could prevent permafrost development in eastern Hudson Bay, Canada [*Menard et al.*, 1998]. Permafrost is always absent under forest or shrubs with snow depth of greater than 80 cm on Hudson Bay coast, Quebec [*Levesque et al.*, 1988], although the role of snow depth in the ground thermal regime depends largely on the particular conditions created by the proximity of Hudson Bay and on the distribution of the surficial deposits.

[45] *Burn* [1998] reported that the impact of changes in seasonal snow cover and soil moisture on permafrost temperatures may surpass the effect of changes in air temperature in northwest North America. However, *Zhang and Stamnes* [1998] demonstrate that the impact of changes in air temperature on permafrost temperatures is greater than that of seasonal snow cover at Barrow, Alaska.

4.3. Sporadic Permafrost

[46] Persistence of permafrost in the sporadic permafrost zone at high latitudes is mainly due to the impact of the minimal snow cover combined with the impact of peatland forms and, in some cases, latent heat in ice-rich permafrost. A peat layer has an opposite impact on the ground thermal regime and permafrost compared with the seasonal snow cover. The peat layer is a layer of a deposit consisting of decayed or partially decayed plant remains with its bulk density less than 400 kg m^{-3}. In summer the peat dries out and becomes an excellent insulator with thermal conductivity ranging from 0.3 to 0.6 W m^{-1} K^{-1}. Autumn rainfall wets the peat, often close to the saturated level. When the nearly saturated peat layer freezes, its thermal conductivity increases dramatically, ranging from 1.1 to 1.6 W m^{-1} K^{-1}. A low thermal conductivity peat layer in summer reduces heat loss from permafrost into the atmosphere, whereas a higher thermal conductivity frozen peat layer in winter, together with thin or no snow cover, increases heat loss from permafrost. As a result, permafrost may exist under the peat layer on the southern margins of permafrost region. For example, *Brown* [1973] reports that in the Yellowknife area the greatest local extent of permafrost is in peatlands and that the mean annual temperature at a depth of 15 m is in the range of about 2°C in granite and −1.0°C in spruce peatland. *Zhang et al.* [1985] report that relic permafrost exists at an elevation of about 2200 m on the south side of the Altai Mountains because of the combined effect of thin snow cover, thick peat layer, and ice-rich permafrost. *Burn* [1998] reports similar results in northwest North America. This combined effect arises because of the marked opposite impact of seasonal snow cover and peat layer on the ground thermal regime [*Williams and Smith*, 1989; *Kudryavtsev*, 1992; *Yershov*, 1998; *Zhou et al.*, 2000]. The persistence of the sporadic and isolated permafrost further complicates interpretation of the responses of permafrost to climate change.

[47] Thin snow cover or absence of snow due to blowing wind is a major factor for permafrost formation or existence in low-latitude and midlatitude high-elevation areas. Low winter snow cover and high winds that remove snow cover above the tree line are major factors that favor the permafrost formation in the mountain ranges of North America [*Harris and Giardino*, 1993]. On the other hand, blowing snow deposited in depressions and on the lee side of slopes may lead to either a warmer or cooler thermal regime of the underlying ground. If the main snowfall occurs in the autumn or early winter with strong winds, the snow insulation effect may prevail and snow cover results in higher ground surface temperatures. Frequently, snow deposited by wind on the lee side of slopes can be a few meters or sometimes even tens of meters in thickness. Thick snow

cover will increase the time required to melt snow, thus reducing the length of time that ground is snow-free in spring or early summer. In this case, snow will cool the ground surface in two ways: by minimizing the effects of incoming solar radiation and by consuming a large amount of energy to melt snow.

[48] The timing of the main snowfall in the autumn vis-à-vis the appearance of cold air greatly affects the results of snow cover impact on the ground thermal regime. *Ives* [1973] reported that the timing of seasonal snow cover in the Colorado Rocky Mountain regions might provide an extremely favorable environment for permafrost formation. The most significant environmental conditions at the Niwot Ridge, Colorado, are relatively low mean annual air temperature (about −1.0°C at 3500 m and −4.0°C at 3750 m) and very high westerly winds in winter that generally maintain a snow-free condition for all but lee slopes above tree line. Heavy snowfall in spring (April–May) results in the greatest extent of snow cover at a time when the midwinter period of most extensive heat loss is well past. The high albedo of snow cover and latent heat consumption due to snowmelt cools the ground surface significantly, a favorable condition for permafrost development in this region.

[49] In mountainous regions the heavy winter snowfall favors the development of glaciers, whereas light winter snowfall correlates with cooling of the ground surface and thus development of permafrost. This implies that periods of expansion of mountain permafrost may not coincide with those of expansion of glaciers [*Harris and Corte*, 1992].

[50] The measurement of the snow-ground interface temperature, or basal temperature, under the dry snow cover by the end of the winter is an effective method of determining permafrost distribution. The method, called the basal temperature of winter snow cover or BTS, was originally developed by *Haeberli* [1973] and applied by many investigators [e.g., *Haeberli and Patzelt*, 1982; *King*, 1983; *Keller*, 1987; *Hoelzle*, 1992; *Keller and Hansueli*, 1993]. The basal temperature of the snow cover is measured before the onset of snowmelt, usually in February and March in the Alps, provided that the minimum snow cover thickness is 80 cm or greater [*King*, 1983]. By the end of the winter, BTS remains nearly constant and mainly controlled by soil heat flux under a relatively thick snow cover (>80 cm). The soil heat flux is strongly influenced by the presence or absence of permafrost. *Hoelzle* [1992] reports that in the northern Alps where the snow cover is more than 100 cm thick by the end of winter, permafrost probably occurs when BTS is less than −3.0°C, permafrost possibly happens when BTS is between −2°C and −3°C; and probably no permafrost exists when BTS is greater than −2.0°C.

4.4. Seasonally Frozen Ground

[51] Compared to permafrost, studies on seasonally frozen ground have received relatively little attention in the past. Seasonal snow cover can lead to higher ground temperatures, reduce ground freezing depth, and increase soil moisture. *Ye* [1964] conducted some preliminary investigations on the impact of the seasonal snow cover on soil temperature and soil moisture in the Urumqi area of the northwestern region of China. He concluded that seasonal snow cover prevents soils from freezing in winter because of its insulation effect and thus increases soil moisture because of meltwater infiltration into soils in spring. Later, *Ma et al.* [1993] demonstrated that snow cover can reduce freezing depth substantially compared to the bare surface (snow was removed after snowfall) in northwest China. *Zhang et al.* [2001] reported that the mean annual air temperature is about −0.5°C at Irkutsk, Russia, where permafrost potentially can develop. Because of the insulating effect of snow cover, mean soil temperature at a depth of 3.2 m is about 3.5°C. Permafrost is absent in the region, and the maximum seasonally frozen thickness is only about 1.5 m.

[52] Recently, comprehensive meteorological measurements, including soil temperature and moisture, were conducted at Rosemount, Minnesota [*Baker and Baker*, 2002]. Air temperature, surface radiation components, and soil temperature at various depths were measured at hourly intervals from October of 1997 through May of 1998. Although the snow depth was measured on an irregular basis over the winter, change of the surface albedo provides an excellent indication of the presence of snow on the ground. Figure 16 demonstrates that the amplitude of the daily soil temperature at a depth of 5 cm was reduced significantly when snow was deposited on the ground surface. Frozen ground initially developed during middle to late November when air temperature was below 0°C (Figure 16a) and snow was deposited on the ground surface (surface albedo was well above 0.4, an indicator of snow on the ground (Figure 16b)). Besides the lower air temperature, fresh snow on the ground may cool the surface because of its high albedo and relatively shallow (a few centimeters) snow cover; therefore the insulating effect was relatively weak. During most of December 1997 and January 1998, snow was absent from the surface, and ground freeze up at depths to 40 cm took place (Figure 16d). During mid-January to mid-February 1998, snow cover was well developed with an average thickness of about 15 to 25 cm, and the depth of frozen ground fluctuated around 40 to 45 cm, even though air temperature was well below 0°C (Figure 16a). Another noteworthy feature is that thawing of the frozen layer was a very rapid process, taking only a few days at the time of spring snowmelt and rising air temperatures.

[53] In some areas, frozen ground may be well developed before snow comes. Using satellite passive microwave remote sensing data, *Zhang et al.* [1999b] demonstrated that, in general, seasonally frozen ground is developed before the establishment of snow cover on a regional scale. As soon as snow cover is established and its thickness increases with time, the frozen ground may start to thaw, and by the end of winter, there may be a very thin frozen layer or no frozen ground at all under the snow cover, especially when snow cover is relatively thick. This is mainly due to the constant heat flux from the deep soil layer to the near surface, and since snow keeps heat from

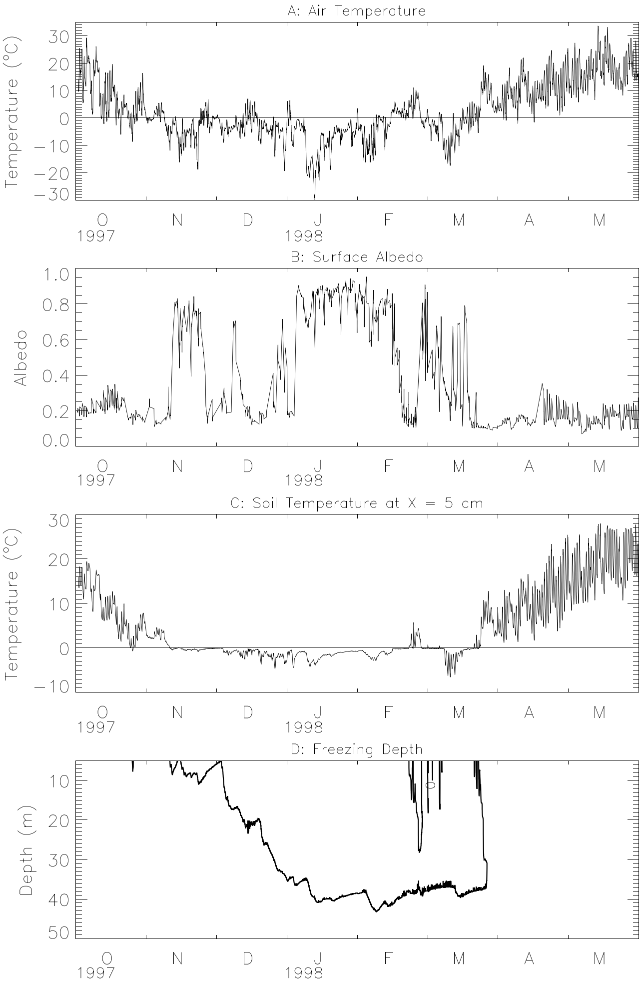

Figure 16. Variations of (a) mean hourly air temperature, (b) surface albedo, (c) soil temperature at 5 cm depth, and (d) freezing depth for winter of 1997–1998 at Rosemount, Minnesota. High surface albedo is referred to as snow presence on the ground surface. Data were obtained from NOAA/Global Energy and Water Experiment (GEWEX) Continental-Scale International Project (available at http://www.joss.ucar.edu/cgi-bin/codiac/dss?18.004).

releasing to the atmosphere, eventually the energy is consumed to thaw the frozen ground.

5. IMPACT OF SNOW COVER ON CLIMATE SIGNAL RECONSTRUCTED FROM DEEP BOREHOLE TEMPERATURE GRADIENTS

[54] It has been recognized that past surface temperature history can be estimated by analyzing the perturbations to the equilibrium geothermal gradient. Indeed, data obtained at various locations around the globe show that temperature gradients are, in fact, disturbed for the first several hundred meters in depth from the ground surface. Surface temperature, reconstructed from deep borehole temperature gradients, shows that some areas have warmed with different magnitudes (up to 4°C) in the past centuries; some show little or no change; some areas show a cooling trend [*Lachenbruch and Marshall*, 1986; *Lachenbruch et al.*, 1988; *Huang et al.*, 1996; *Pollack et al.*, 1998; *Pollack and Huang*, 2000; *Huang et al.*, 2000; *Harris and Chapman*, 2001; *Beltrami*, 2002a, 2002b; *Osterkamp*, 2003]. The magnitude of warming reconstructed from deep borehole temperature gradients contrasts with some proxy-based estimates of surface air temperature histories that suggest lesser amounts of warming during the same period [*Mann et al.*, 1998, 1999; *Crowly and Lowery*, 2000; *Briffa et al.*, 2001; *Esper et al.*, 2002; *Briffa and Osborn*, 2002; *Mann and Jones*, 2003]. There may be many explanations for these differences [*Smerdon et al.*, 2003]; coupling or decoupling between air temperature and ground surface temperature may be one of the major reasons [*Zhang and Osterkamp*, 1993; *Zhang et al.*, 1997; *Osterkamp and Romanovsky*, 1999; *Harris and Chapman*, 2001; *Beltrami*, 2002a, 2002b; *Mann and Schmidt*, 2003; *Mann et al.*, 2003; *Schmidt and Mann*, 2004; *Pollack and Smerdon*, 2004]. Influence of seasonal snow cover on ground temperatures certainly modifies the climate signals reconstructed from deep borehole temperature gradients.

[55] Seasonal snow cover experiences not only seasonal variations but also interannual and historical changes. The interannual and historical changes in seasonal snow cover can also have a significant influence on the ground thermal regime and permafrost development. Figure 17 shows variations of mean annual air temperature and annual maximum thickness of seasonal snow cover at Prudhoe Bay Deadhorse Airport and permafrost temperatures at depths of 20 and 30 m at West Dock for the period from 1983 through 1993 [*Zhang et al.*, 1997; *Romanovsky et al.*, 2003]. During the period from 1983 through 1985, mean annual air temperature was relatively low with very thin snow cover (less than 10 cm), which is a very favorable condition for permafrost cooling (Figure 17a). Both mean annual air temperature and annual maximum thickness of snow cover started to increase in 1986, reaching their maximum values over the entire recorded period in 1989. After 1989 both mean annual air temperature and snow thickness decreased from their peak values. The permafrost temperatures at depths of 20 and 30 m generally followed the variations of mean annual air temperature (MAAT) and annual maximum snow thickness (Figure 17b). However, there was a lag of about 2 to 3 years between MAAT, annual maximum snow thickness, and permafrost temperatures. In 1991, mean annual air temperature was one of the two lowest values in the period of record, which is a favorable condition for cooling permafrost, while in the same year annual maximum snow thickness was one of the two deepest values (about 25 cm), which is a favorable condition for warming permafrost. The combined effect on permafrost temperature of the lowest mean annual air temperature and deepest snow on the ground in 1991 over the period of record was that it seemed to cool the permafrost with a 2–3 year lag of permafrost temperatures. Using a validated one-dimensional numerical model, *Zhang and Stamnes* [1998] show that, in terms of percentage change from their means, the impact of changes in the seasonal snow cover has a lesser effect on the thermal regimes of the active layer and

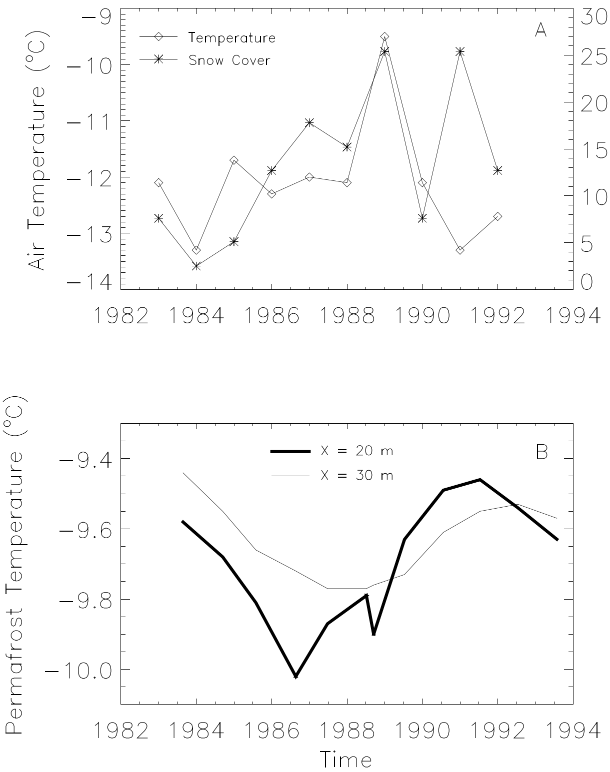

Figure 17. Variations of (a) the mean annual air temperature and maximum thickness of the seasonal snow cover at Prudhoe Bay and (b) permafrost temperatures at depths of 20 and 30 m at West Dock, Prudhoe Bay, Alaska, from 1983 through 1993 [*Zhang et al.*, 1997]. Copyright John Wiley and Sons Ltd. Reproduced with permission.

permafrost than the impact of changes in MAAT in northern Alaska. However, further field measurements are definitely needed to enhance our understanding of the relationships among MAAT, annual maximum snow thickness, and permafrost temperatures in this region.

[56] It is well established that permafrost surface temperature increased about 2°C to 4°C in some areas across the North Slope of Alaska, and the warming started in the 1910s to 1950s, as revealed by analytical modeling [*Lachenbruch and Marshall*, 1986; *Lachenbruch et al.*, 1988]. At some locations, there was little or no change or even a cooling trend. However, there is no evidence of a 2°C to 4°C warming of air temperature in the Barrow record, in the North American quadrant of the Arctic, or in the Arctic as a whole during the same period of the permafrost warming [*Hansen and Lebdeff*, 1987; *Serreze et al.*, 2000]. In this case, variations in the magnitude of the permafrost surface warming might be related to local factors, such as soil type, vegetation, microrelief, soil moisture condition, and seasonal snow cover. During this time, there was a trend with greater snowfall during cold years and smaller snowfall during warmer years at Barrow and Barter Island, Alaska, except for a period from the mid-1960s to the mid-1970s at Barter Island (Figure 18) [*Zhang and Osterkamp*, 1993]. This implies that thicker snow cover during colder years would prevent the ground from cooling, while for thinner snow cover during warmer years it would be relatively easier for heat transfer from the atmosphere to warm the ground. As a result, the effect of changes in air temperature on permafrost surface temperature may have been modified substantially by changes in the seasonal snow cover in the region [*Zhang and Osterkamp*, 1993].

[57] One of the major obstacles to understanding the linkage between the soil thermal regime and climatic change is the lack of widespread long-term observations of soil temperatures and related climate variables. Such measurements were conducted at Irkutsk, Russia, from 1898 through 1995 [*Gilichinsky et al.*, 1998; *Zhang et al.*, 2001]. On the basis of long-term measurements at Irkutsk, Russia, changes in both mean annual air temperature and soil temperature at 40 cm depth were about the same magnitude (2.0°C to 2.5°C), but the patterns of change were substantially different [*Zhang et al.*, 2001]. Mean annual air temperatures increased slightly until the 1960s (Figure 19a), whereas mean annual soil temperature increased steadily throughout the entire period (Figure 19b). Changes in air temperature alone at Irkutsk, Russia, cannot account for the changes in soil temperatures from 1898 through 1995. Furthermore, soil temperature actually decreased during summer months by up to 4°C, while air temperature increased slightly (Figure 20). This cooling in

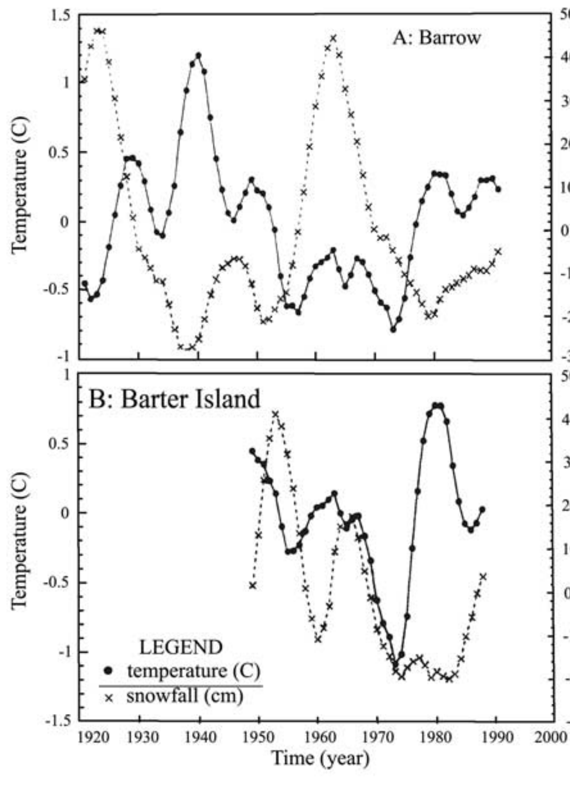

Figure 18. Deviations of the mean annual air temperature and snowfall from their long-term means at (a) Barrow and (b) Barter Island, smoothed by a low-pass filter with a cut-off frequency of 0.091 yr^{-1} (approximately 11-year running mean equivalent) [*Zhang and Osterkamp*, 1993]. Reprinted with permission from South China University of Technology Press.

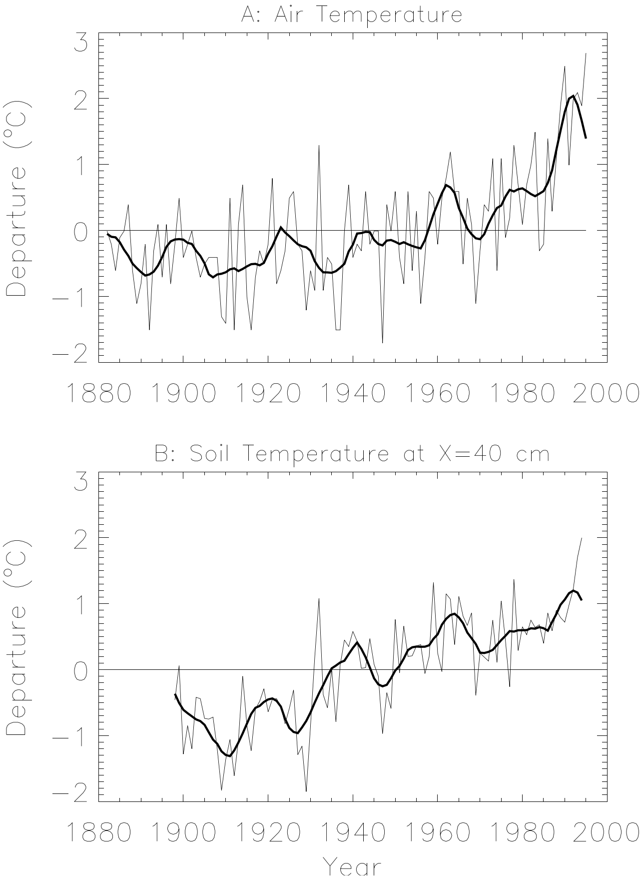

Figure 19. Departures of (a) mean monthly air temperature for the period from 1882 through 1995 and (b) mean annual soil temperature at 40 cm depth from 1898 through 1995 from their long-term means at Irkutsk, Russia.

the soil may be explained by changes in rainfall and hence in soil moisture during summer because of the effect of the soil moisture feedback mechanism. Whereas mean annual air temperature increased about 4°C to 6°C during winter, soil temperature increased by up to 9°C (Figure 20). The increase in snowfall during early winter (October and November) and early snowmelt in spring may play some role in the increase of soil temperatures through the effects of insulation and albedo changes. According to *Stone et al.* [2002] the melt date in northern Alaska has advanced by about 8 days since the mid-1960s. Modeling results indicate that changes in the timing of spring snowmelt by 10 days may result in an increase of mean annual ground surface temperature by 0.2°C [*Ling and Zhang*, 2003]. Obviously, changes in spring snowmelt date alone cannot account for the observed soil temperature increase.

6. SUMMARY AND DISCUSSIONS

[58] Seasonal snow cover during the cold season of the annual temperature cycle is widespread in the Northern Hemisphere and plays an important role in influencing the ground thermal regime. Snow surface has a high albedo that leads to a reduction in absorbed solar energy and a lowering of snow surface temperature. Snow also has a relatively higher thermal emissivity that causes an increase in the outgoing longwave radiation, thus cooling the snow surface. At the same time, snow surface has a relatively higher absorptivity, especially under cloudy sky conditions, that absorbs the atmospheric downwelling longwave radiation, hence warming the snow surface. Snow has an extremely low thermal conductivity and thus acts as an excellent insulator between the atmosphere and the land surface. Snowmelt is an energy sink because of the fusion of latent heat. The net effect of snow cover on ground thermal regime and its magnitude depends on the timing, duration, and accumulation and melting processes; the thickness, density and structure of the seasonal snow cover; interactions of micrometeorological conditions, local microrelief, and vegetation; and geographical location.

[59] Early establishment of the seasonal snow cover in autumn may be one of the best favorable conditions for snow insulating effect, while later snowfall in the late winter or early spring may have a cooling effect on the ground thermal regime. Snow thickness is another important parameter that has a significant impact on the ground thermal regime. Changes in snow thickness have a great impact on soil temperatures when snow thickness is less than 50 cm or so. Snow density has a direct impact on thermal conductivity of snow cover and thus its insulating effect.

[60] In continuous permafrost regions, seasonal snow cover can lead to an increase of the mean annual ground temperatures and permafrost surface temperature by several degrees, while in discontinuous and sporadic permafrost regions, the absence of seasonal snow cover may be a key factor for permafrost development. In seasonally frozen ground regions, snow cover can substantially reduce the seasonal freezing depth. However, influence of seasonal snow cover on seasonally frozen ground has received relatively little attention, and further study is much needed.

[61] Interannual variations of seasonal snow cover conditions (timing, duration, density and structure, and thickness) have significant impact on long-term ground temperature variability in cold seasons/cold regions. Ground surface temperature histories reconstructed from deep borehole temperature gradients have been widely used as evidence of paleoclimate change. The linkage between the reconstructed ground surface temperature and the air temperature has not been fully understood. Certainly, changes in air temperature alone cannot account for the changes in ground temperature. Changes in snow cover conditions significantly modify the impact of air temperature on ground thermal regime and might have contributed considerably to the ground surface temperature increase in the past centuries. Better knowledge is needed to further understand the impact of air temperature and snow cover, as well as other parameters such as precipitation, vegetation, and soil moisture on ground thermal regime.

[62] Both in situ snowpack manipulations and laboratory experiments have demonstrated that winter heterotrophic activity increases following spring snowmelt and soil thaw because of an increase in labile carbon substrates [*Schimel et al.*, 2001]. Because snow cover insulates soil from cold

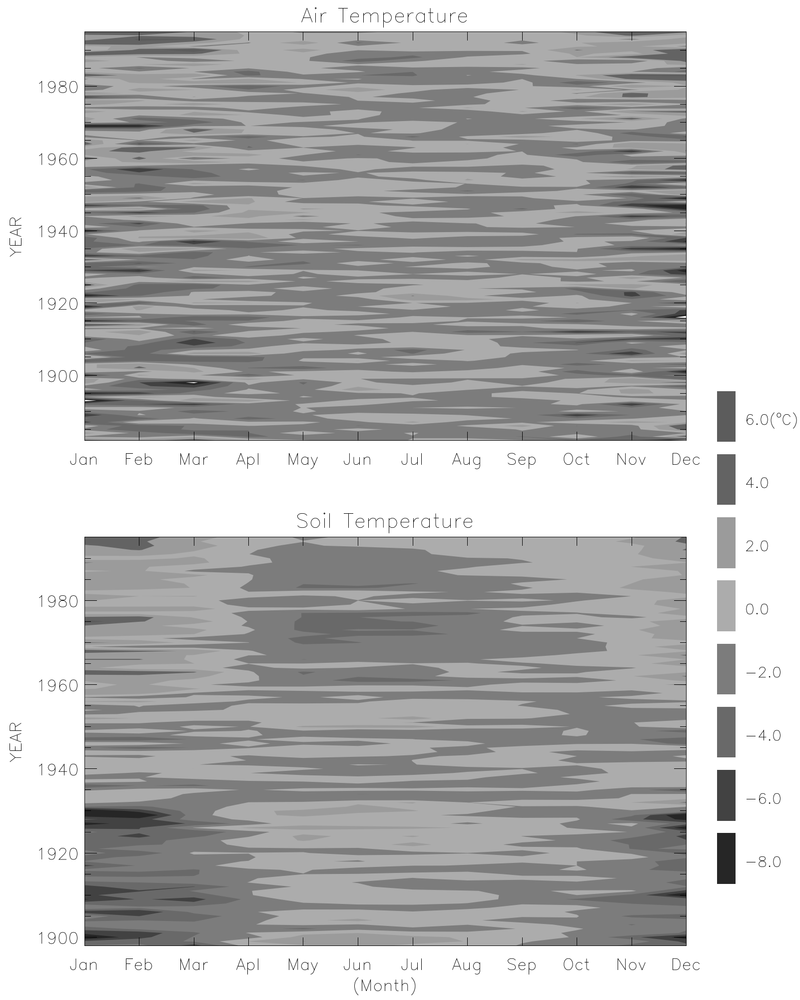

Figure 20. Departures of the (top) mean monthly air temperature from its long-term mean (1882–1995) and (bottom) mean monthly soil temperature from its long-term mean (1898–1995) at Irkutsk, Russia. See color version of this figure at back of this issue.

atmospheric temperatures over winter months, shallower snow cover results in colder soils and more severe soil frost. If snow cover remains shallow, soil remains frozen and microbial activity is suppressed throughout the winter. In contrast, if a deeper snowpack develops later in the season, previously frozen soil may thaw, and microbial activity and heterotrophic respiration may be greatly enhanced. Conversely, similar levels of soil frost may enhance microbial activity and soil decomposition processes at one location, while reducing it at another. Consequently, the timing of snowpack development appears to have significant effects on subniveal biogeochemical activity and subsequent fluxes of carbon and nitrogen. Exchange of carbon, methane, and other gases between the atmosphere and the land surface may be minimal when the soil is frozen, but it is accentuated in the early spring following thaw. *Skogland et al.* [1988] called this release a "respiratory burst" [also see *Sommerfeld et al.*, 1993].

[63] The hydrology community has long recognized the importance of the freeze/thaw cycle. Freezing of soil moisture reduces hydraulic conductivity leading to either more runoff due to decreased infiltration or higher soil moisture content due to restricted drainage. The existence of a thin frozen layer near the surface essentially decouples soil moisture exchange between the atmosphere and the deeper soils underneath the frozen soil. Knowing whether the soil is frozen or not is important in predicting surface runoff and spring soil moisture reserve [*Willis et al.*, 1961; *Cary et al.*, 1978]. Knowing the soil freeze/thaw status under the snowpack is particularly important. If soil is concrete frozen under snow cover in spring, snow meltwater will completely contribute to runoff and potentially produce high risk of floods; whereas if soil is partly frozen or unfrozen under snow cover, snow meltwater will penetrate into the soil and reduce runoff, potentially lowering the possibility of spring snow meltwater floods. Our recent study indicated that the near-surface soils generally froze before snow covered the land surface during the 1997/1998 winter over the contiguous United States [*Zhang and Armstrong*, 2001]. The number of days for which soil was frozen before snow had accumulated on the ground surface ranged from 1 to 125 days, with an average value of about 4 days.

[64] Recent studies indicate that snow area extent has generally decreased, with a substantial seasonal and inter-

annual variability, in the United States and in the Northern Hemisphere as a whole during the past 20 years [*Armstrong and Brodzik*, 2001]. Studies also indicate that the decrease in annual snow cover extent was accompanied by an increase in air temperature by almost 1°C in North America [*Leathers and Robinson*, 1993]. The overall decline in snow cover area extent implies that more bare land surface is directly exposed to the atmosphere, thus potentially increasing the areal extent and thickness of the seasonally frozen soils. On the other hand, increase in air temperature during years with less snow may result in less frozen soil area extent. A key question then is the response of the seasonally frozen soils, such as the timing, duration, area extent, and thickness, to changes in air temperature and seasonal snow cover with respect to climatic change under the global warming scenarios and the potential feedback to the climatic system.

[65] Climate change will likely result in warmer winters leading to less snowfall. These changes may lead to decreases in soil temperatures and increases in soil freezing because of lack of an insulating snow cover (i.e., colder soils in a warmer world) and changes in soil water dynamics during the important snowmelt period. Results from snow manipulation experiments in the northeastern United States indicate that even mild winters with low snowfall resulted in increased soil freezing [*Groffman et al.*, 2001; *Handy et al.*, 2001]. These results suggest that a climate shift toward less snowfall or a shorter duration of snow on the ground will produce increases in soil freezing in northern latitudes. The feedback of increased soil freezing to the climate system is poorly understood and needs further investigation.

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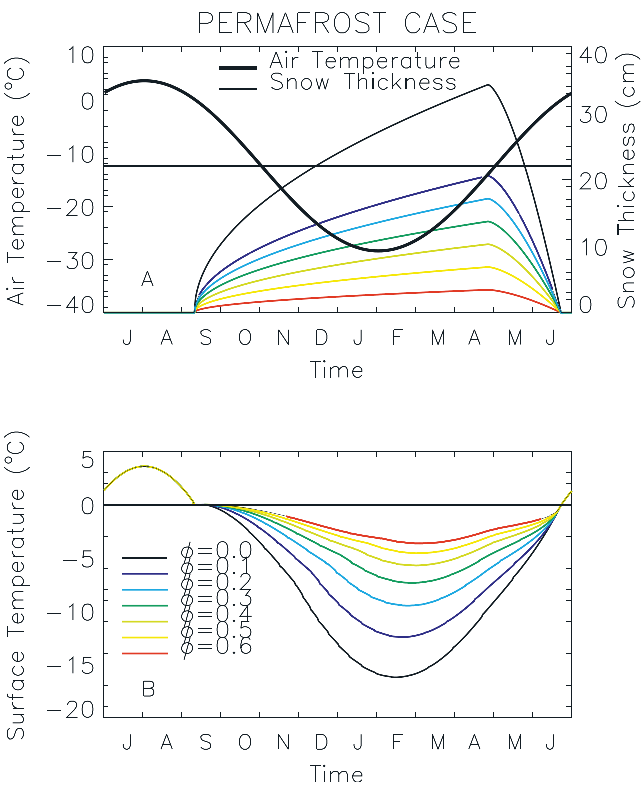

Figure 13. Impact of changes in the depth hoar fraction of the seasonal snow cover on ground surface temperature. (a) Input air temperature, mean annual air temperature (solid horizontal line), and snow and depth hoar thickness for the case along the Alaskan Arctic coast. (b) Output ground surface temperatures with variations of the depth hoar fraction in the seasonal snow cover. Line types in Figures 13a and 13b are consistent. From *Zhang et al.* [1996c].

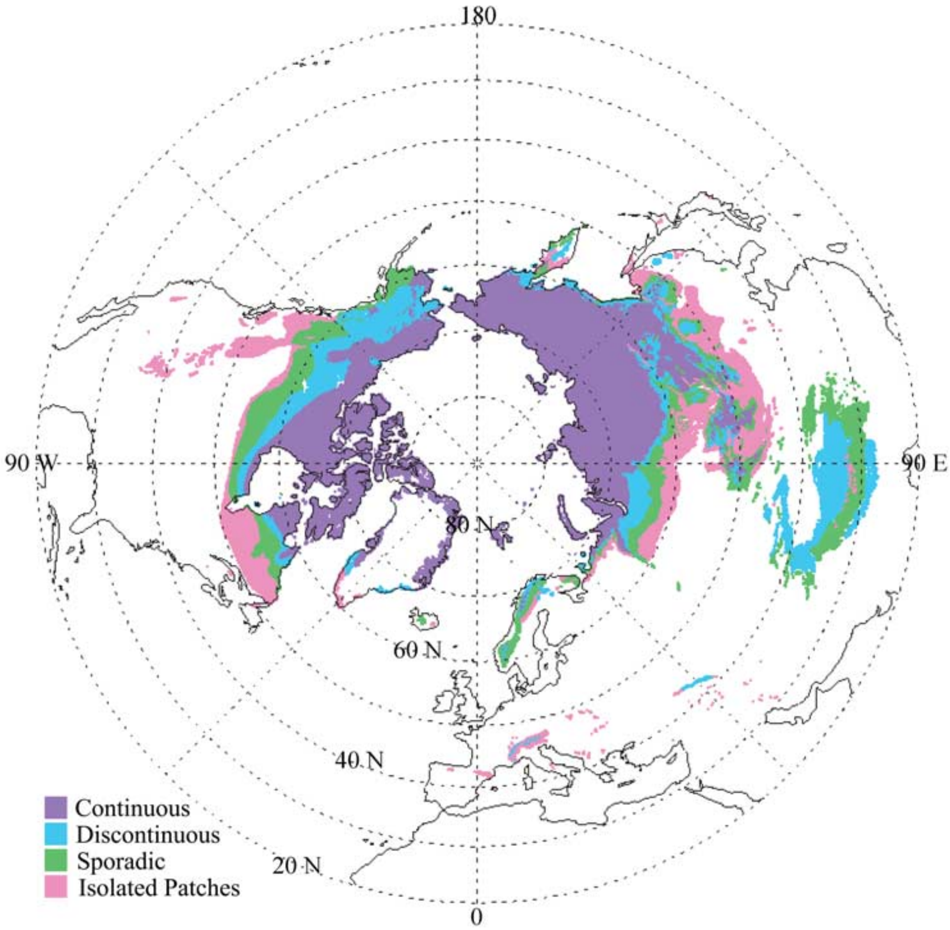

Figure 14. Permafrost distribution in the Northern Hemisphere. Data were obtained by *Brown et al.* [1997] and simplified by *Zhang et al.* [1999a]. Reprinted with permission of V. H. Winston and Son, Inc.

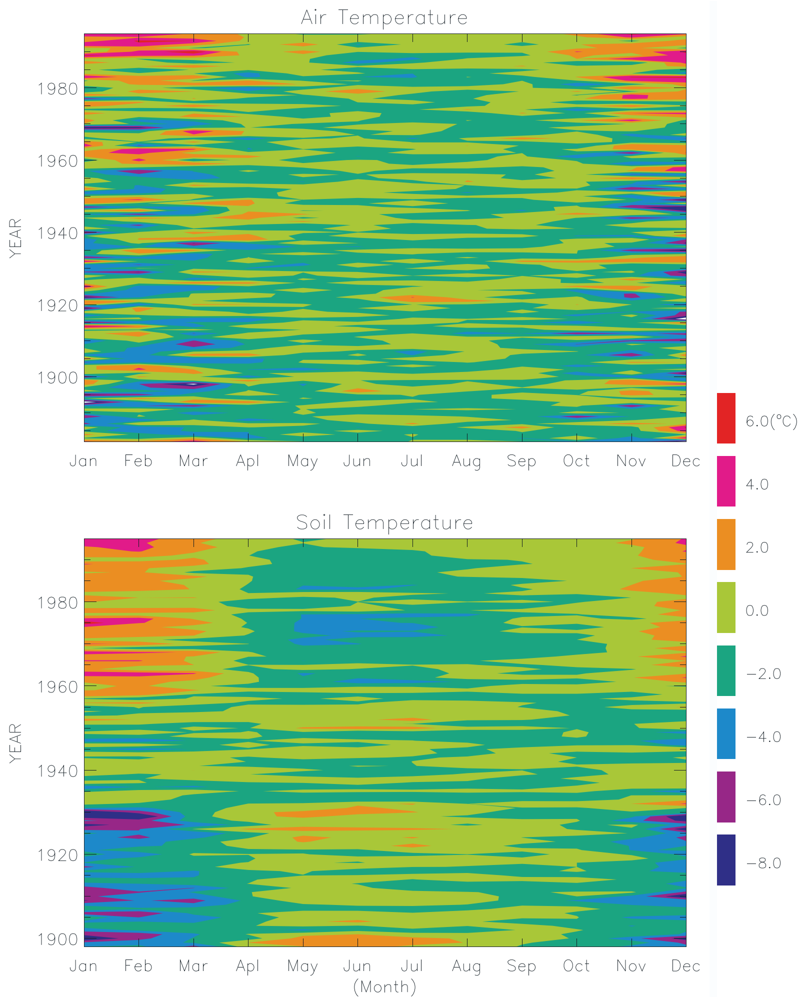

Figure 20. Departures of the (top) mean monthly air temperature from its long-term mean (1882–1995) and (bottom) mean monthly soil temperature from its long-term mean (1898–1995) at Irkutsk, Russia.