AN AMPLIFIED SIGNAL OF CLIMATIC CHANGE IN SOIL TEMPERATURES DURING THE LAST CENTURY AT IRKUTSK, RUSSIA

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Abstract. Climatic changes at the Earth's surface propagate slowly downward into the ground and modify the ambient ground thermal regime. However, causes of soil temperature changes in the upper few meters are not well documented. One major obstacle to understanding the linkage between the soil thermal regime and climatic change is the lack of long-term observations of soil temperatures and related climatic variables. Such measurements were made throughout the former Soviet Union with some records beginning at the end of the 19th century. In this paper, we use records from Irkutsk, Russia, to demonstrate how the soil temperature responded to climatic changes over the last century. Both air temperature and precipitation at Irkutsk increased from the late 1890s to the 1990s. Changes in air temperature mainly occurred in winter, while changes in precipitation happened mainly during summer. There was an anti-correlation between mean annual air temperature and annual total precipitation, i.e., more (less) precipitation during cold (warm) years. There were no significant trends of changes in the first day of snow on the ground in autumn, but snow steadily disappeared earlier in spring, resulting in a reduction of the snow cover duration. A grass-covered soil experiences seasonal freezing for more than nine months each year and the long-term average maximum depth of seasonally frozen soils was about 177 cm with a range from 91 cm to 260 cm. The relatively lower soil temperature at shallow depths appears to represent the so-called 'thermal offset' in seasonally frozen soils. Changes in mean annual air temperature and soil temperature at 40 cm depth were about the same magnitude (2.0 °C to 2.5 °C) over the common period of record, but the patterns of change were substantially different. Mean annual air temperature increased slightly until the 1960s, while mean annual soil temperature increased steadily throughout the entire period. This leads to the conclusion that changes in air temperature alone cannot explain the changes in soil temperatures at this station. Soil temperature actually decreased during summer months by up to 4 °C, while air temperature increased slightly. This cooling in the soil may be explained by changes in rainfall and hence soil moisture during summer due to the effect of a soil moisture feedback mechanism. While air temperature increased about 4 °C to 6 °C during winter, soil temperature increased by up to 9 °C. An increase in snowfall during early winter (October and November) and early snowmelt in spring may play a major role in the increase of soil temperatures through the effects of insulation and albedo changes. Due to its relatively higher thermal conductivity compared to unfrozen soils, seasonally frozen ground may enhance the soil cooling, especially in autumn and winter when thermal gradient is negative.

Climatic Change **49:** 41–76, 2001.

1. Introduction

The thermal regime of the soils is a combined product of energy and mass exchange between the atmosphere and the land surface. Soil temperature is a sensitive climate indicator and integrator and plays an important role in all the physical, biological and microbiological processes occurring in the soil. The need to understand the thermal conditions of soils and their relation to the environmental conditions have long been a preoccupation of scientists. Such interest has been mostly motivated by applications to agriculture (e.g., De Vries, 1963; Sharratt et al., 1992) and, in the northern latitudes, to the study of the evolution of permafrost and its stability in the context of development and extraction of natural resources (Kudryavtsev, 1954; Lachenbruch et al., 1962; Brown, 1982; Lachenbruch and Marshall, 1986; Osterkamp and Lachenbruch, 1990; Nelson et al., 1993; Pavlov, 1996). Current interests concern reconstructing the past climate from deep borehole temperature profiles (Lachenbruch and Marshall, 1986; Taylor, 1991; Osterkamp et al., 1994; Pollack et al., 1998) and climate model predictions of high latitude warming (Anisimov and Nelson, 1996; Takata and Kimoto, 1998; Li et al., 1998).

Air temperature variations propagating through the ground interface, with intervening vegetation and seasonal snow cover, are recorded in the subsurface as perturbations on the long-term equilibrium geothermal gradient. It has long been recognized that past surface temperature history can be estimated by analyzing the perturbations to the equilibrium geothermal gradient (e.g., Lane, 1923; Hotchkiss and Ingersoll, 1934; Birch, 1948). Indeed, data obtained at various locations around the globe show that temperature gradients are in fact disturbed for the first several hundred meters. After careful examination and analysis to eliminate the non-climatic causes of the nonlinearities in the deep soil temperature profiles, it has been established that surface temperature has increased up to 4 °C over a time scale ranging from decades to centuries (Lachenbruch and Marshall, 1986; Beltrami and Mareschal, 1991; Taylor, 1991; Osterkamp et al., 1994; Gosnold et al., 1997; Huang et al., 1997; Pollack et al., 1998). This climatic information obtained from deep borehole temperature profiles has been used as a robust complement to the existing paleoclimatic databases (e.g., Pollack and Chapman, 1993; Beltrami and Taylor, 1995; Harris and Chapman, 1997; Overpeck et al., 1997; Serreze et al., 2000) as well as providing the long-term records of temperature change needed for the validation of General Circulation Models (GCMs) for future climate estimates.

Because of the general concern about the potential response of the ground thermal regime, especially the seasonal freeze/thaw and permafrost in cold regions, to predicted global warming, there is a need for better understanding of the processes and the environmental factors governing the energy and mass exchange between the atmosphere and the land surface (Smith, 1975; Goodrich, 1982; Outcalt et al., 1990; Kane et al., 1991; Nelson et al., 1993; Zhang and Osterkamp, 1993; Pavlov, 1994, 1996; Zhang et al., 1997a). Understanding of the linkage between the subsurface soil temperature and its relation to environmental factors remains a critical issue for the combined analysis of deep soil temperature profiles and proxy data and for testing climatic inferences.

One major obstacle to understanding the relationship between the soil thermal regime and its environmental conditions is the lack of long-term observations of soil temperature and related climatic variables. Such measurements were made throughout the former Soviet Union, with some records beginning in the last century and many others beginning in the 1930s or 1950s. In this paper, we will examine the long-term records of soil temperature at various depths, air temperature, precipitation, snowfall, and snow thickness data at Irkutsk Observatory (52° N, 104° E, and 467 m a.s.l.), Russia. This station was selected for the present detailed case study because of the comprehensive records. Our aim is to improve understanding of soil temperature climatology and its variations during the period of record and to investigate how soil temperature responds to the related climatic variables.

2. Data Sources and Methods

The basic data used in this study include: mean monthly air temperature; mean monthly soil temperatures at depths of 40 cm, 80 cm, 160 cm, and 240 cm; the first and last day of soil freezing at the ground surface, 40 cm, 80 cm, and 160 cm; seasonal maximum freezing depth; monthly total precipitation; and the timing and thickness of the seasonal snow cover. The derived parameters include mean annual air, ground surface, and soil temperatures at various depths, annual total precipitation, duration of seasonal snow cover and soil freezing at depth of 40 cm, 80 cm, and 160 cm, snow cover index, snowfall (water equivalent), annual freezing and thawing indices of air and soil temperatures at 40 cm depth.

Meteorological observations began in 1882 and soil temperature measurements started in 1898 at Irkutsk. Both standard meteorological and soil temperature observations have continued without interruption to the present. Data are available to use through 1995. Over the entire period of field measurements, there was no site movement at this station. The city is located in the southeastern part of the undulating Irkutsk Basin. It is on the boundary of the forest-steppe which is to the northwest of the station. The observatory park is sited in the southern district of Irkutsk on a 35–40 m terrace above the Angara River which is 1 km to the west. A reservoir (1.5–4.0 km wide) was created by a dam on the Angara River in 1956, 3 km to the southeast of the Observatory. Grey podzolized forest soils overlie friable loamy terrace sediments. The soil profile in the meteorological enclosure is summarized in Table I.

The station 'horizon' from east, through south, to west is between 15° and 20° elevation from trees and buildings; a building having a 23° elevation is to the northwest. In the south, the horizon is below 10°. There are two observational sites for soil temperature measurements: one is the standard Russian meteorological station

TABLE I
Soil type at the Irkutsk Observatory

Depth (m)	Description of soil types
0.0–0.4	A medium-grained dark loam
0.4–1.4	A brown fine-grained loam
1.4–4.0	A grey structureless loam with sand veins in the lowest meter

with grass surface, another has a bare soil surface after removal of the vegetation. In this study, data from the standard meteorological station with a grass surface were used.

It is generally true that precipitation falls as rain or snow when mean monthly air temperature is well above or below 0 °C during summer or winter months, respectively. Lauscher (1976) reported that solid precipitation as a percentage of total precipitation varies linearly with mean monthly air temperature (MMAT), from 100% when MMAT < -10 °C and 0% when MMAT > 10 °C in the northern hemisphere. This relationship may overestimate snowfall or solid precipitation. Ross and Walsh (1986) adopt the assumption that during the transition months the fraction of precipitation falling as snow varies linearly from 100% for $MMAT < -5$ °C to 0% for MMAT of 3 °C. Serreze et al. (1998) used this method to estimate precipitation and snowfall in the high Arctic and obtained good results. This method was used here to calculate annual total rainfall and snowfall.

The freezing and thawing indices are defined as the cumulative number of degree-days below and above 0 °C for a given time period (Permafrost Subcommittee, 1988). In this study, the annual freezing and thawing indices were calculated using monthly mean values of air and soil temperatures at 40 cm depth. A comparative analysis indicates that the error is less than 5% for annual air temperature freezing and thawing indices calculated from daily and monthly mean air temperatures at Barrow, Alaska (Zhang et al., 1996), although the error could be greater in lower latitudes where the transition seasons can last longer.

Seasonal snow cover is one of the primary factors influencing the ground thermal regime in cold regions. The net effect of the seasonal snow cover on the ground thermal regime and its magnitude depends upon the timing, duration, thickness, density and structure of the snowpack and its interactions with other micrometeorological conditions, local microrelief, and vegetation (Goodrich, 1982; Zhang et al., 1997a). It is very difficult, and sometimes misleading, to evaluate the effect of snow on the ground thermal regime spatially and temporally using a single snow cover parameter such as snow cover duration or its maximum thickness. In this paper, we introduce the concept of the snow cover index, SCI, the integration of the snow cover thickness over time:

$$\mathrm{SCI} = \int_{t_1}^{t_2} H_s(t)dt\ , \tag{1}$$

where t_1 and t_2 are the first and last day of snow on the ground, respectively; H_s is snow thickness at time t. t_1 and t_2 are determined when snow thickness at the ground surface is at or greater than 2.0 cm. The duration of snow cover is defined as the period from t_1 through t_2. From Equation (1), SCI accounts for the duration and thickness of seasonal snow cover over time. In practice, the calculation of SCI should be:

$$\mathrm{SCI} = \sum_{i=1}^{N} H_i \Delta t_i\ , \tag{2}$$

where N is the total number of measurements and Δt_i is the time interval between two consecutive measurements in day. The snow cover index is a measure of combined duration and thickness of the seasonal snow cover during winter months. The unit of SCI is in cm-day.

Mean monthly air temperature, monthly total precipitation, and snow cover data were provided by the EOSDIS Distributed Active Archive Center (DAAC) at the National Snow and Ice Data Center, University of Colorado. Soil temperatures were provided by the Institute of Soil Science, Russian Academy of Sciences, Pushchino, through the Russian Hydrometeorological Institute, Obninsk, and archived at the National Snow and Ice Data Center, University of Colorado.

The procedures used to measure soil temperature at stations of the Hydrometeorology Service of the former Soviet Union are described by Gilichinsky et al. (1998) and are detailed in instruction manuals of the State Committee of the U.S.S.R. for Hydrometeorology and Environmental Control (1985) and earlier issues between 1946 and 1969. A series of reports provide data from 1891 at the stations of the hydrometeorological network (State Committee of the U.S.S.R., 1960–1964; 1970–1978 and 1966–1990). Information on the station site characteristics, surroundings and history are published by the State Committee of the U.S.S.R. (1964–1972). A summary of essential information follows. Temperature measurements are made by an alcohol or mercury thermometer at a grass covered soil surface in the warm season and the snow surface in winter. Also in the warm season, soil temperatures are determined by bent-stem thermometers at 0.05, 0.10, 0.15 and 0.20 m depth. Because of this seasonal limitation, those data are not used in this study. Temperatures at 0.2–3.2 m depth are recorded by extraction thermometer installed in an ebonite pipe beneath a grass covered plot, with the natural snow cover during the snow season. The standard depths are: 0.2, 0.4, 0.8, 1.6, and 3.2 m, sometimes with additional measurements at 0.6, 1.2, and 2.4 m. The 0.4 m level was selected here because it provides the full record. At this depth, observations were made at 0700, 1300, and 2100 hours before 1936, at 0100, 0700, 1300, and 1900 hours from 1936 to 1965, and at 0000 and every 3 hours Moscow standard time (UTM + 3 hours) since 1966. Since 1959, observations have been rounded to the nearest whole degree Celsius. Only monthly data were acquired by the National Snow and Ice Data Center, because funds were inadequate to support digitizing of the voluminous day/hour records. The snow depth measurements at Russian stations are made once daily using the average of the depths at three fixed stakes in the meteorological enclosure.

3. Climatic Conditions

3.1. AIR TEMPERATURE

The long-term mean annual air temperature at Irkutsk from 1882 through 1995 was –0.5 ± 0.9 °C, with a mean annual minimum of –2.2 °C in 1947 and a mean annual maximum of 2.2 °C in 1995. However, for the period from 1882 to the early 1960s, mean annual air temperature was generally below its long-term mean value, increasing slightly with a rate of about 0.0073 °C/yr (Figure 1). For the period from the 1960s to the mid-1990s, mean annual air temperature was mostly above its long-term mean value and has risen about 2.0 °C, with a rate of 0.0800 °C/yr, more than 10 times greater than the rate from the early 1880s to the mid-1960s (Figure 1). Over the period of record, the mean annual air temperature at Irkutsk, Russia, increased about 2.5 °C, primarily during the last three decades. These results are consistent with trends obtained from other parts of Russia (Pavlov, 1996).

The long-term annual average freezing and thawing indices of air temperature are 2258 ± 283 °C-day and 2090 ± 99 °C-day, respectively (Figure 2). Although the long-term mean values of the freezing and thawing indices are of the same magnitude, inter-annual variations of the freezing index are significantly greater than that of thawing index. For example, the standard deviation of the freezing index is about 13% of its long-term mean, while the standard deviation of thawing index is less than 5%. The freezing index decreased slightly from 1882 to the early 1960s with a rate of –1.78 °C-day/yr, while the rate of decrease was about –21.38 °C-day/yr from the mid-1960s to the early 1990s. The pattern of decrease in the freezing index (Figure 2a) is in agreement with the increase of mean annual air temperature (Figure 1). This demonstrates that changes in the mean annual air temperature at Irkutsk are mainly controlled by changes in the freezing index, or changes in subzero air temperature during winter months.

3.2. PRECIPITATION

The long-term average annual total precipitation was about 448 ± 104 mm, with a minimum of 209 mm in 1884 and a maximum of 883 mm in 1938 (Figure 3). Precipitation at Irkutsk mainly falls as rain during summer months with a long-term average of 367 ± 98 mm, accounting for about 82% of the annual total precipitation (Figure 4c).

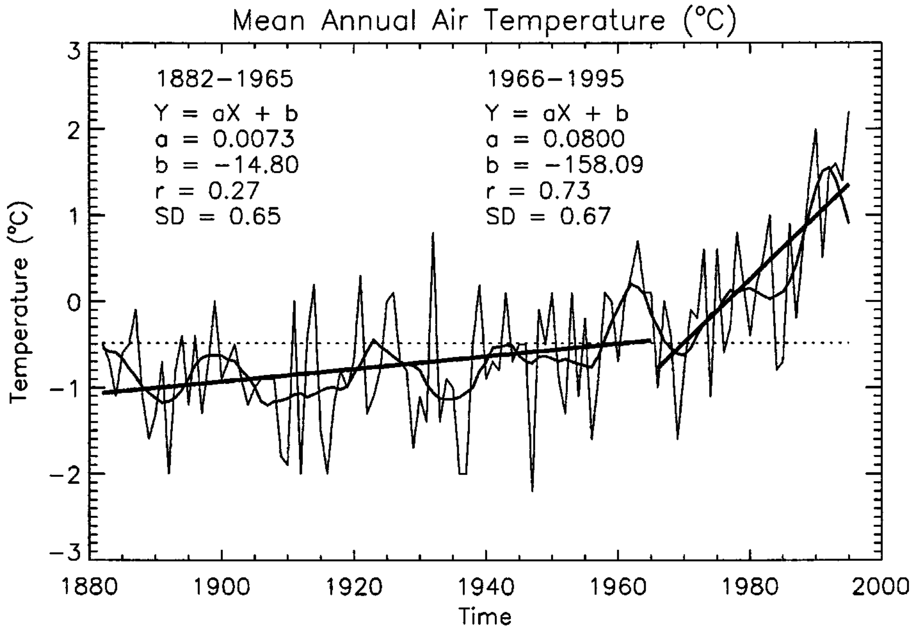


Figure 1. Variations of the mean annual air temperature at Irkutsk, Russia, from 1882 through 1995. The long-term mean annual air temperature (dashed line) is about –0.5 °C. The thin solid line represents the mean annual air temperature; the mid-thick solid line indicates a smoothing trend using a low-pass filter with a cut-off frequency of 0.091; the thick solid line is a linear regression with the parameters shown on the upper-left corner, where *Y* stands for air temperature in °C and *X* for time in calendar years, *a* and *b* are the correlation constants, *r* is the correlation coefficient, and SD is the standard deviation of the measured mean annual air temperature from the linear regression line.

Historically, there was a trend of increasing precipitation over time with an increasing rate of about 1.2 mm/yr (Figure 3). Annual total precipitation was generally below its long-term average for the period from the early 1880s to 1930. It was well above its long-term average from the early 1930s to mid-1970s. Precipitation increased from 300 mm in the early 1880s to 450 mm around 1905, then decreased to 370 mm around 1925, and increased rapidly to 550 mm over the next 10 to 15 years. Annual total precipitation fluctuated around 500 mm from the early 1940s to 1970s, decreasing to around 400 mm in 1980, and recovering slightly to about 470 mm in the mid-1990s. Since approximately 82% of the annual total precipitation fell as rain, annual total rainfall had a similar pattern of variation as the annual total precipitation (Figure 4a). The long-term annual total snowfall (water equivalent) was 81 $\pm$ 26 mm. Snowfall increased at an annual rate of about 0.19 mm/yr from about 70 mm at the beginning to about 92 mm at the end of the record (Figure 4b). Further analysis indicates that snowfall increased from 40 mm in the 1880s to 100 mm in the early 1900s, then to 70 mm in the early 1930s, and increased again to 110 mm in the early 1960s; since then, snowfall decreased to 70 mm in the mid-1990s. Although there were large fluctuations in rainfall as a

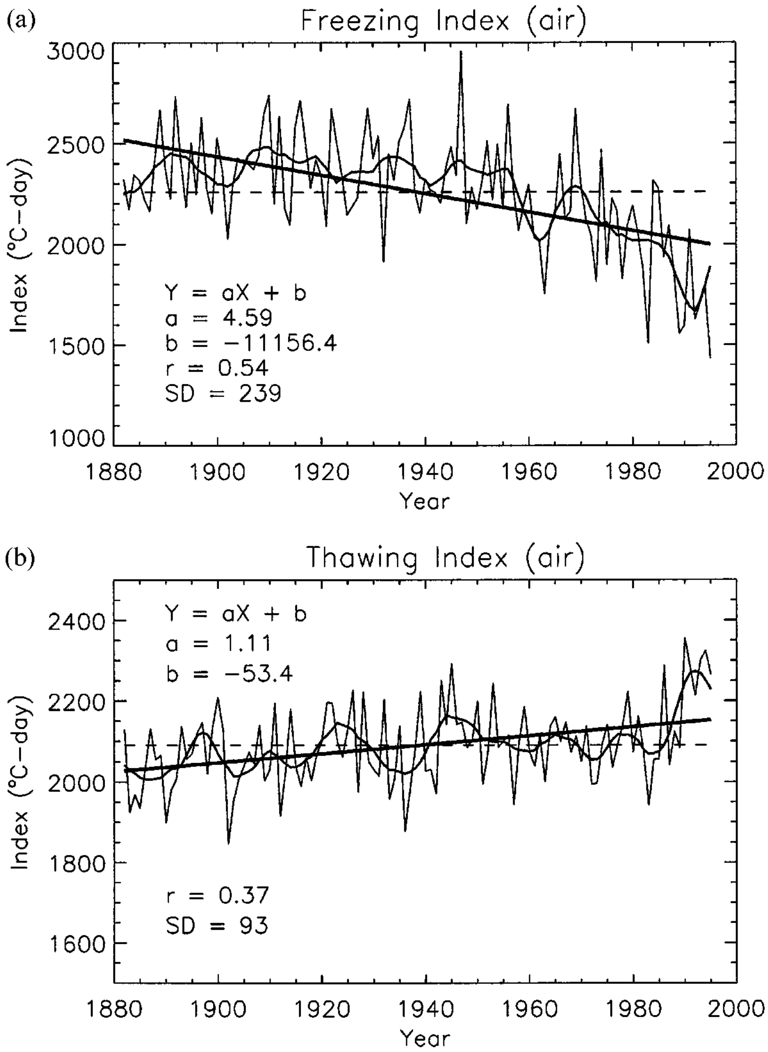


Figure 2. Same as Figure 1 for freezing index (a) and thawing index (b) of air temperature. The long-term mean values for the freezing and thawing indices are 2258 °C-day and 2090 °C-day, respectively.

fraction of the annual total precipitation, with extremes varying from greater than 93% in 1971 to less than 66% in 1909, there were no significant trends of changes in the rainfall fraction (Figure 4c).

3.3. SEASONAL SNOW COVER

The timing, duration, and thickness of the seasonal snow cover changed significantly during the period 1935–1985 (Figure 5). On average, the first day of snow on the ground occurred around the day of the year (DOY) 295 (Figure 5a). Generally,

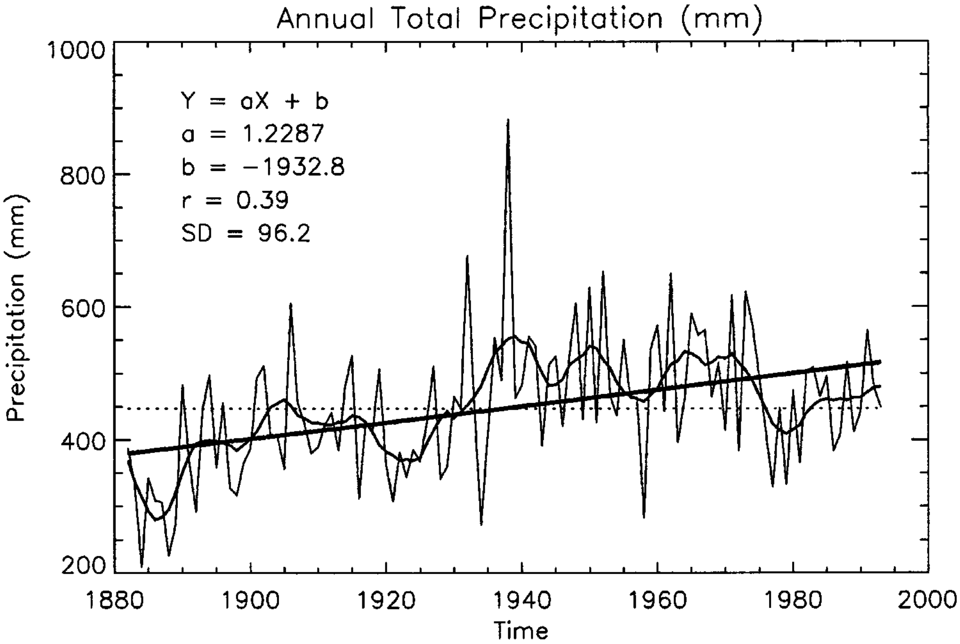


Figure 3. Same as Figure 1 for annual total precipitation from 1882 through 1993. The long-term annual total precipitation is about 448 mm/yr.

snow cover establishment started in mid- to late October from the late 1930s to around 1960, then began in early October around 1965, reverting to late October in the mid-1980s. During the extreme years, the first day of snow on the ground could vary over two months, varying from early September in 1968 to late November in 1971. However, there was no statistically significant trend of changes in the first day of snow on the ground at Irkutsk. The last day of snow on the ground occurred, on average, at the beginning of April (DOY 95) (Figure 5c). There is a general trend for snow cover to disappear earlier over the period of record, from late April in the late 1930s to late March in the 1980s. Further analysis indicates that the date of snow disappearance changed from DOY 110 around 1940 to DOY 80 in 1965 and then became 5 to 10 days later in the 1980s. The duration of the seasonal snow cover, on average, was about 170 days (Figure 5e). The duration became shorter over the period of record, ranging from six months in the 1940s to about five months in 1980s. During the extreme years, snow cover duration ranges from 201 days in 1939 to 128 days in 1971.

The average snow thickness over the entire winter was approximately 20 cm over the period of record (Figure 5b), but generally decreased from about 25 cm in the late 1930s to 15 cm around 1980. The average maximum snow thickness by the end of winter was about 40 cm with extremes varying from a low of 21 cm in 1971 to a high of 103 cm in 1953 (Figure 5d). The overall trend of the maximum snow thickness decreased with time. Maximum snow thickness was about 45 cm at the

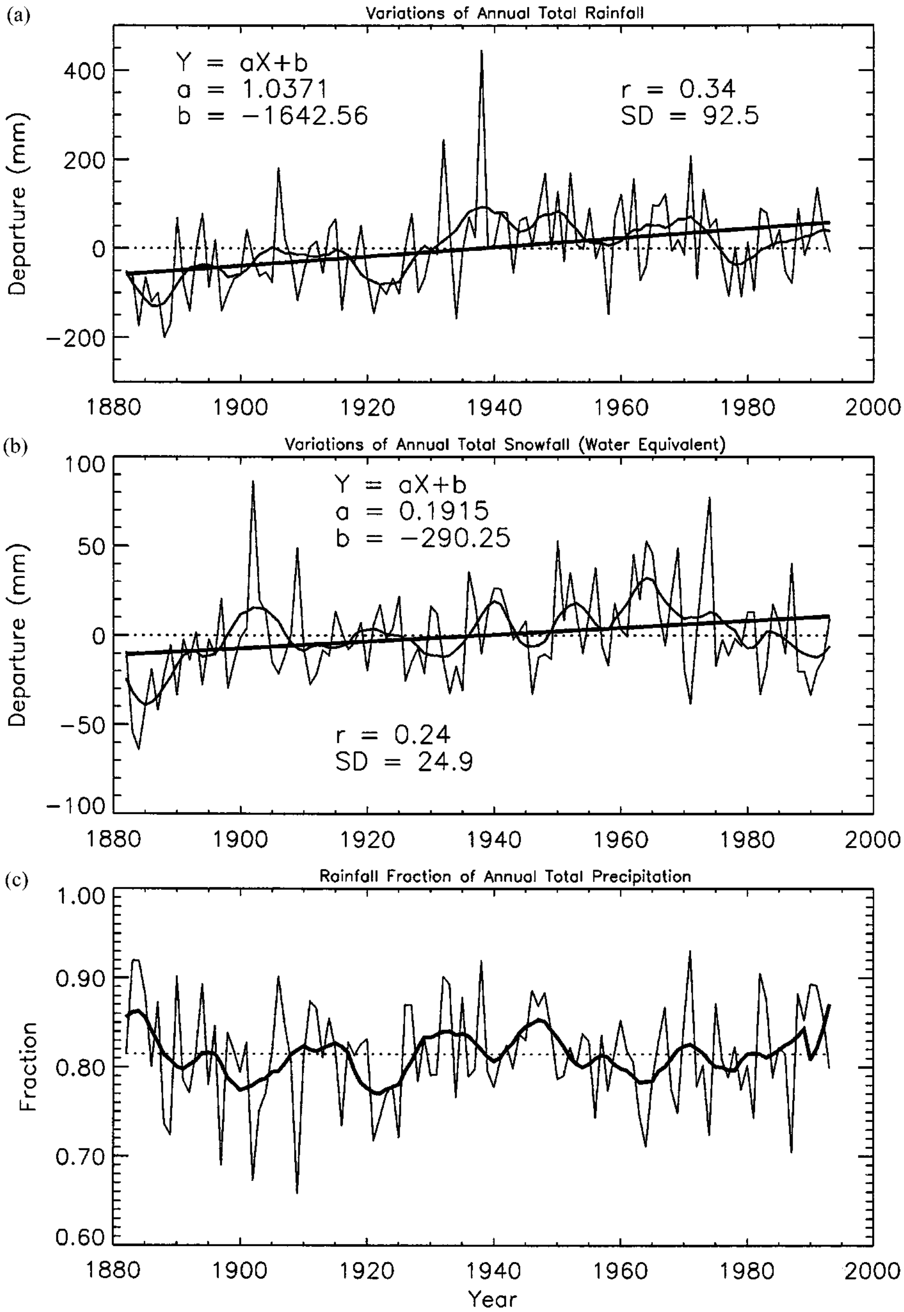


Figure 4. Same as Figure 1 for annual total rainfall (a), annual total snowfall (b), and the rainfall fraction of the annual total precipitation (c). The long-term mean annual rainfall and snow-fall (water equivalent) are about 368 mm/yr and 89 mm/yr, respectively.

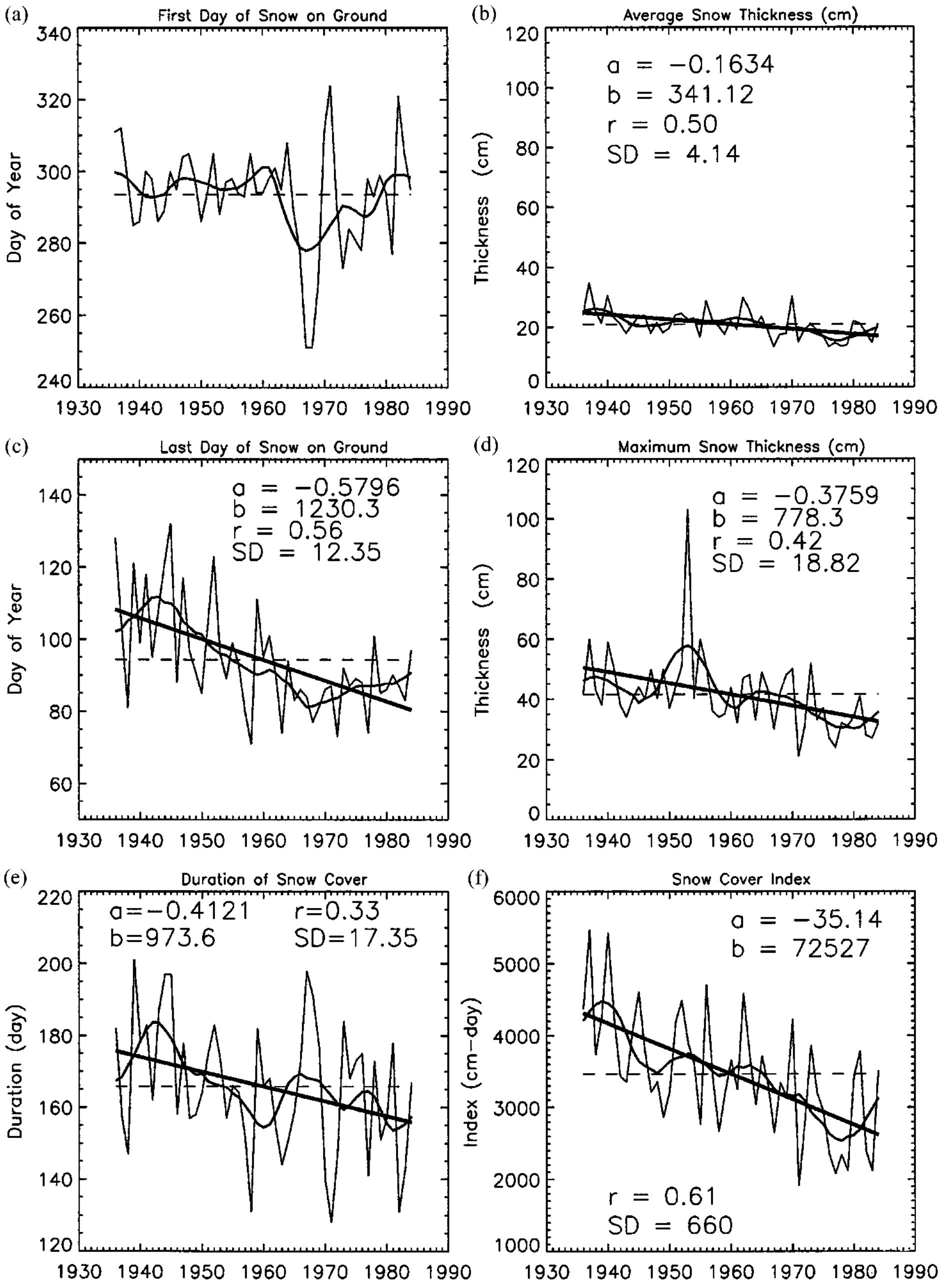


Figure 5. Variations of the snow cover parameters from 1935 through 1985 at Irkutsk, Russia. (a) first day of snow on the ground; (b) average snow thickness over the entire snow season; (c) last day of snow on the ground; (d) maximum snow thickness by the end of winter; (e) duration of seasonal snow cover; and (f) snow cover index. The long-dashed line indicates the average values for each variable; the thin solid line indicates annual values; the mid-thick solid line indicates smoothed values using a low-pass filter with a cut-off frequency of 0.091; the thick solid line indicates linear regression line. The parameters a, b, r and SD are the same as in Figure 1, except for designated variables.

beginning of the record, decreasing to about 35 cm in the mid-1940s. It increased to over 55 cm in the beginning of the 1950s and then decreased steadily to about 30 cm around 1980.

The snow cover index (SCI) (Figure 5f) decreased steadily from 4400 cm-day in the late 1930s to 2600 cm-day in the 1980s. It decreased sharply from 4500 cm-day in the late 1930s to its long-term average value of 3465 cm-day by the end of the 1940s. During the 1940s and early 1950s the SCI was relatively flat with the values slightly above its long-term average value. The SCI then experienced another sharp decrease from about 3500 cm-day in the mid-1960s to about 2500 cm-day in the late 1970s before recovering to about 3000 cm-day in the mid-1980s.

Overall, both mean annual air temperature and annual total precipitation increased over the period of record. Changes in air temperature mainly occurred in winter months while changes in precipitation mainly in summer months at Irkutsk, Russia. There was a general anti-correlation between mean annual air temperature and annual total precipitation, i.e., more precipitation during cold years and less precipitation during warmer years, except in the early 1960s when there was more precipitation with relatively high air temperature. Therefore, wetter summers were related to the colder winters in the region. Seasonal snow cover, one of the major factors that influence the ground thermal regime, changed substantially. Although there are no significant trends of changes in the first day of snow on the ground, snow disappeared earlier in spring, resulting in a reduction of the duration of the seasonal snow cover. The snow cover index, a combined measure of the duration and thickness of seasonal snow cover, decreased substantially due to the early disappearance of snow and the decrease in snow thickness.

4. Soil Thermal Regimes

4.1. SOIL TEMPERATURES

The near-surface soil temperature regime and its change with time are a result of climatic conditions and their change. The long-term mean annual soil temperature from 1898 through 1994 at Irkutsk was about $2.8 \pm 0.8\,°C$ at 40 cm depth below the ground surface, and increase slightly with depths, to $2.9 \pm 0.8\,°C$ at 80 cm, $3.0 \pm 0.7\,°C$ at 160 cm, and $3.2 \pm 0.5\,°C$ at 240 cm. The long-term mean soil temperature at 40 cm depths was about $0.4\,°C$ lower than that at 240 cm depth. The lower soil temperature at shallower depth is due to the impact of seasonal freezing and thawing processes or 'thermal offset' effect (Kudryavtzev et al., 1974; Goodrich, 1978; Burn and Smith, 1988; Romanovsky and Osterkamp, 1995). Detailed discussions of the impact of seasonal freezing and thawing on soil temperatures will be provided in the following sections.

Figure 6 demonstrates variations of annual soil temperature departures from their long-term means at depths of 40 cm, 80 cm, 160 cm, and 240 cm. As expected, the patterns of change are consistent at all depths with the magnitude tending to diminish exponentially with depths. This indicates that the soil temperature is mainly controlled by the surface boundary conditions and there is no internal heat source within the depth of interest. This gives us confidence that the measured soil temperature at any depth is of good quality and can be used to investigate response to climatic conditions on the surface.

There was a general increase in soil temperatures with time at all depths with a rate of change of 0.0223 °C/yr at 40 cm, 0.0199 °C/yr at 80 cm, 0.0165 °C/yr at 160 cm, and 0.0122 °C/yr at 240 cm (Figure 6). The mean annual soil temperature increased about 2.14 °C at depth of 40 cm, 1.92 °C at 80 cm, 1.58 °C at 160 cm, and 1.17 °C at 240 cm, over the period of record. The superposed fluctuations have magnitudes of less than 1 °C.

4.2. SEASONALLY FROZEN SOILS

Seasonally frozen soil is a product of heat exchange between the atmosphere and the land surface. Because of a relatively lower air temperature (MAAT $= -0.5$ °C, Figure 1) at Irkutsk, seasonally frozen soil is well developed and one of the main natural phenomena in this region. The timing, duration, and thickness of seasonally frozen soils are basic parameters to understand the seasonal soil freezing and thawing in cold regions. The first and last day of soil freezing (at any depth) is defined as the first and last day when soil temperature is at or below 0 °C in autumn and spring, respectively. The duration of soil freezing is determined as a difference between the last and first day of soil freezing. The annual maximum freezing depth was determined as the maximum depth of 0 °C isotherm by the end of winter.

For the period from 1947 through 1959, the first day of a grass-covered surface soil freezing ranged from 13 August to 16 September with an average of 26 August. The last day of surface soil freezing ranged from 25 May to 30 June with an average of 11 June. The duration of seasonally frozen soils at Irkutsk ranged from 251 days in 1959 to 312 days in 1949, with an average of about 289 days over 13 years. Thus, a grass-covered soil experiences seasonal freezing for more than nine and a half months annually at Irkutsk, Russia. However, it should be understood that the surface soil is not necessarily continuously frozen for the period from the first to last day of surface soil freezing.

Figure 7 indicates the first day (a), the last day (b), and duration (c) of soil freezing at 40 cm depth and annual maximum freezing depth (d) for the period from 1899 through 1970 at Irkutsk, Russia. The first day of soil freezing at 40 cm depth ranged from 20 October 1913 to 10 December 1932 with an average around 11 November and an inter-annual variation of up to 51 days. The first day of soil freezing at 40 cm depth was almost linearly delayed, on average, about 10 days from around early November in 1900 to mid-November in 1930, then advanced by about a week until the early 1950s, and finally was delayed by about 10 days again until the late 1960s. This pattern of change in the first day of soil freezing at 40 cm depth is closely related to the variation of mean monthly air temperatures

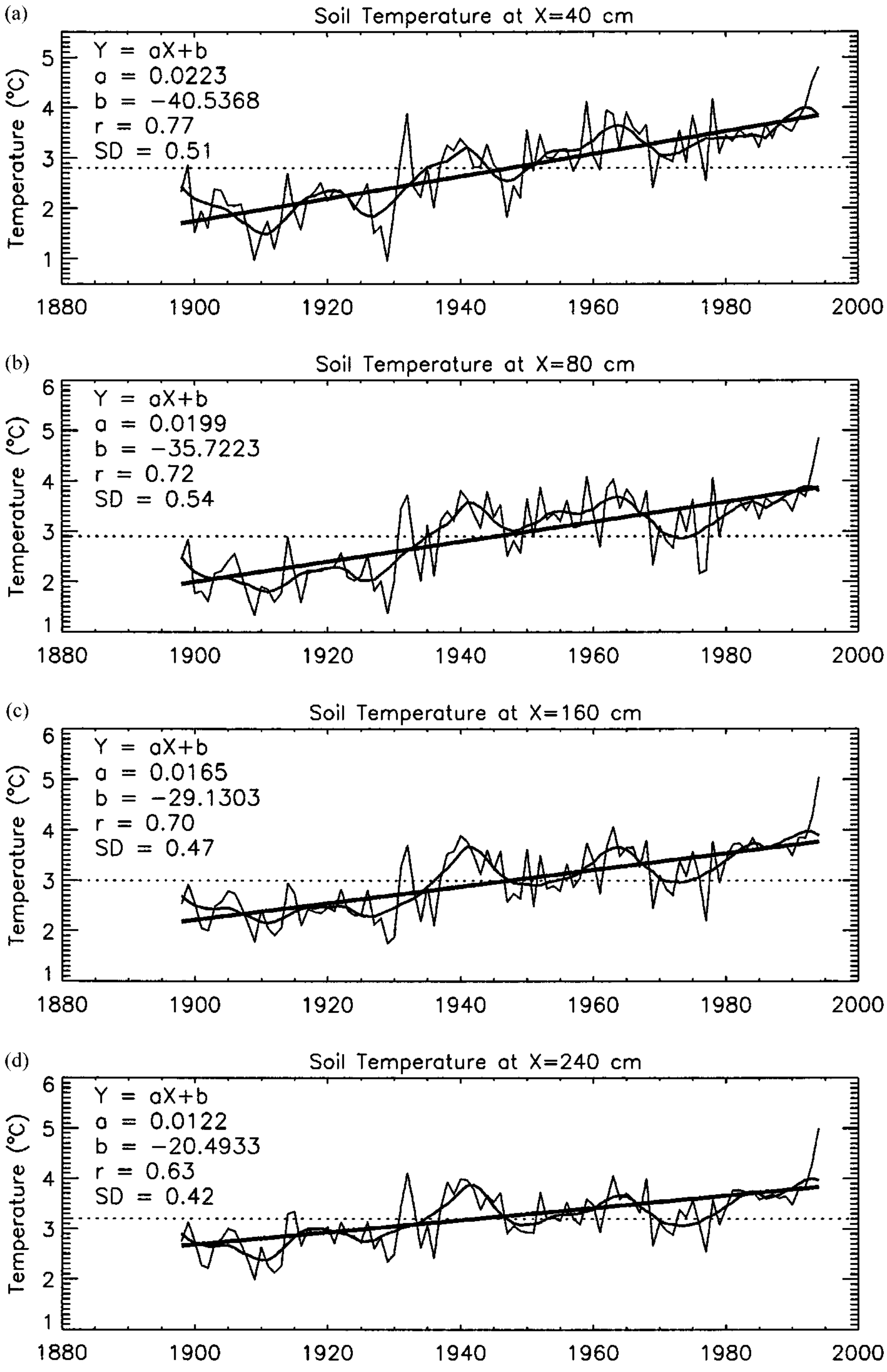


Figure 6. Same as Figure 1 except for soil temperatures at depths of 40 cm (a), 80 cm (b); 160 cm (c); and 240 cm (d). The long-term mean soil temperatures (1898 through 1995) were 2.8 °C at 40 cm; 2.9 °C at 80 cm; 3.0 °C at 160 cm; and 3.2 °C at 240 cm.

in November (Figure 10k), i.e., the cold periods in the early 1900s and around the 1950s were responsible for the early freezing, while the relatively warm periods around the early 1930s and late 1960s were responsible for the late soil freezing. The early snow on the ground in the late 1960s (Figure 5a) might also have some impact on the late soil freezing at 40 cm depth due to the insulating influence of snow cover on heat exchange between the atmosphere and the ground surface.

The last day of soil freezing at 40 cm depth ranged from 10 April 1922 to 13 May 1970 with an average around 16 April and inter-annual variations of up to 33 days. For the period from 1900 through the mid-1910s, the smoothed value (thick solid line in Figure 7b) of the last day of soil freezing at 40 cm depth was just slightly above its long-term average (dashed-line in Figure 7b) with significant inter-annual fluctuations. Then, it advanced about two weeks until the mid-1920s, delayed about one week until the mid-1930s, and advanced again for more than a week until the early 1940s. Finally, the last day of soil freezing at 40 cm depth was delayed for about three weeks from the mid-1940s through the late 1960s. The last day of soil freezing at any depth is controlled by the combined effect of air temperature, snow cover thickness and last day of snow on the ground, soil temperature prior to thawing of the frozen soils, and soil moisture conditions in spring.

The duration of soil freezing at 40 cm depth ranged from 139 days in 1939 to 189 days in 1913 with an average of 166 days or about five and half months annually over the period of record. The inter-annual variation of the duration was up to 50 days. The annual maximum freezing depth varied from 91 cm in 1963 to 260 cm in 1901 with an average value of about 177 cm. In general, the annual maximum freezing depth was well below and above its long-term average for the periods from 1900 to 1930 and from 1940 to 1970, respectively, with a sharp decrease from 1930 through 1940. At a given location, changes in winter air temperature or freezing index, seasonal snow cover, and soil moisture conditions prior to soil freezing may contribute most to changes in the annual maximum freezing depth although there are many other factors which may also influence the seasonal freezing and thawing processes. At this stage, there is not enough information on snow cover and soil moisture to explain the variations of the annual maximum freezing depth over the entire period of record. However, the sharp decrease in the annual maximum freezing depth from the late 1920s to 1940 (Figure 7d) might be due to the increase in snowfall (Figure 4b) and slight decrease in freezing index of air temperature (Figure 2a) over the same period. The decrease in the freezing depth during the mid-1960s (Figure 7d) might be due to the decrease in freezing index (Figure 2a), increase in snowfall (Figure 4b), and early snowfall (Figure 5a) over the same period. A noteworthy feature is that the duration of soil freezing at 40 cm depth (Figure 7c) was essentially decoupled from the seasonal maximum freezing depth (Figure 7d). For example, the duration of soil freezing at 40 cm depth was above its long-term average value for the periods from 1900 to 1920 and from the early 1950s to 1970 (Figure 7c), respectively, which correspond to both below

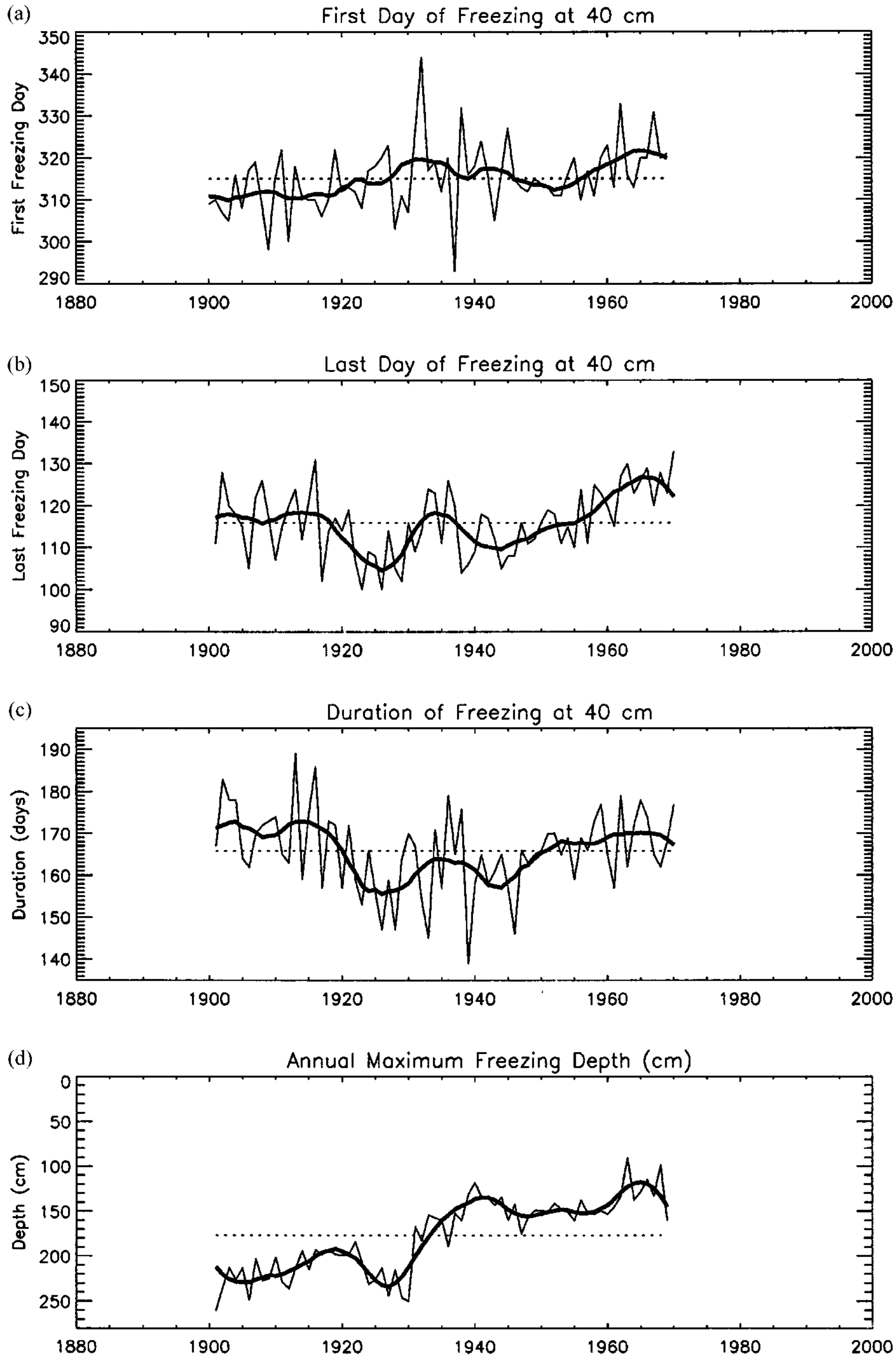


Figure 7. Variations of the first day of soil freezing (a), last day of soil freezing (b), and duration of soil freezing (c) at 40 cm depth, and annual maximum freezing depth (d) for the period from 1900 through 1970 at Irkutsk, Russia. The thin solid line represents annual values; the thick solid line represents smoothed value using a low-pass filter with a cut-off frequency of 0.091; the dashed line represents the long-term mean values.

and above the long-term average value of the seasonal maximum freezing depth (Figure 7d). For the period from 1920 through 1950, the duration was generally below its long-term average value, while the seasonal maximum freezing depth experienced substantial decrease over the same period.

5. Relations of Soil Temperatures to Climatic Conditions

In this section, we will investigate the relationship between soil temperature and climatic conditions over the past century at Irkutsk, Russia. Ideally, mean monthly ground surface temperature should be used for this analysis. However, the surface temperature was measured at the ground surface when there was no snow on the ground and on the snow surface when snow cover existed. The mean monthly ground surface temperature was also not continuous over the period of record. Instead, mean monthly soil temperature at 40 cm depth is used for this analysis. Considering the extreme continental climate (large diurnal air temperature variation) and sandy soils (implying higher thermal conductivity) at Irkutsk, the depth of daily soil temperature variation corresponding to the diurnal air temperature change should be well below 40 cm depth. The presence of seasonal snow cover during winter months may delay the response of soil temperature to changes in air temperature. However, soil thermal conductivity would increase substantially due to soil freezing compared with thawed soils. Thus, the lag of changes in soil temperature at 40 cm depth relative to changes in air temperature should be one to two days, at most, during winter months. On a monthly basis, changes in soil temperature at 40 cm depth should directly respond to changes in air temperature, surface conditions, snow cover, and soil moisture. On this basis, we can justifiably use soil temperature at 40 cm depth to investigate the response of soil temperatures to changes in air temperature, precipitation, seasonal snow cover, and seasonally frozen soils.

5.1. CLIMATOLOGY

Figure 8 illustrates the climatology of air temperature (a), precipitation (b), and soil temperature (c) at 40 cm depth at Irkutsk over the period of record. Soil temperatures were much higher than air depth temperature during winter months (October through March), and lower during the summer months (April through August). Overall, the long-term mean annual soil temperature was about 3.3 °C higher than the long-term mean annual air temperature. The annual amplitude of 40 cm soil temperature was about 3.3 °C, substantially less than the 19 °C annual amplitude of air temperature.

The magnitude of difference between soil temperature at 40 cm depth and air temperature during winter months has exceeded 12 °C, while during summer months the difference was generally within –5 °C (Figure 8d). This asymmetric

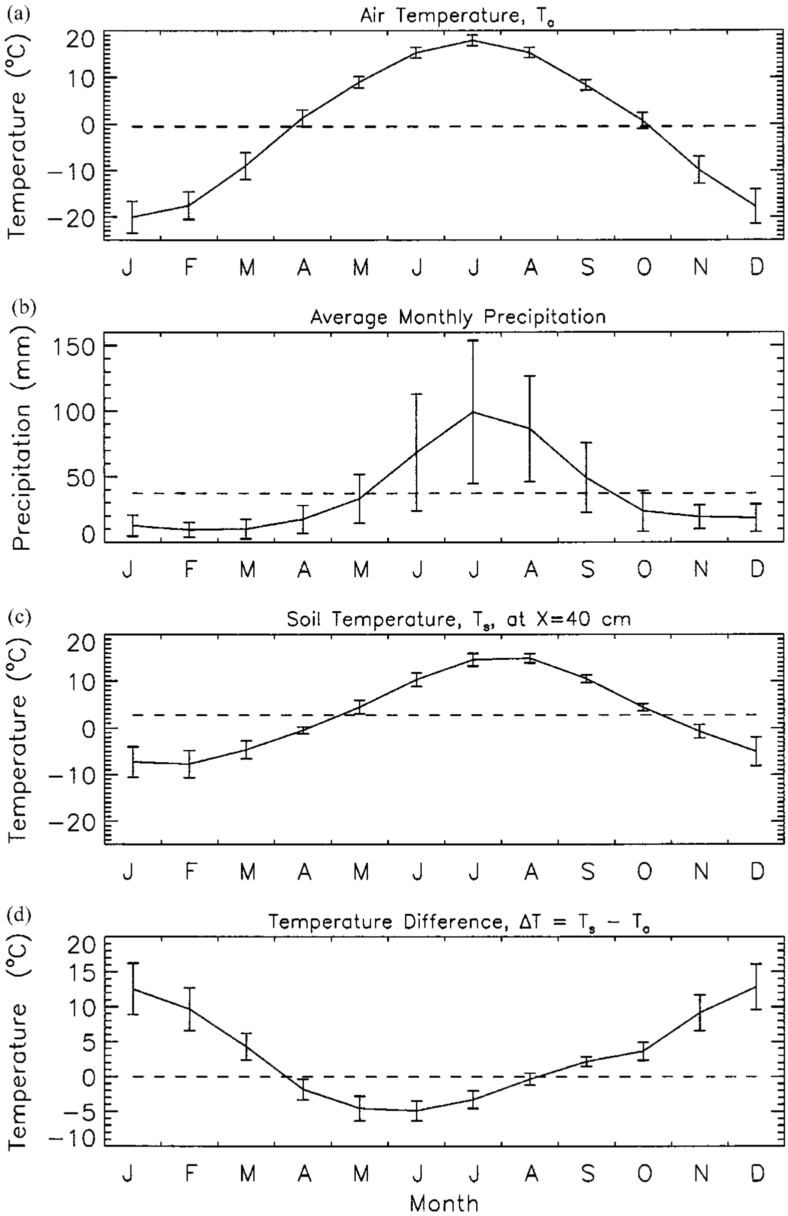


Figure 8. The long-term mean monthly air temperature (a), total precipitation (b), soil temperature at 40 cm depth (c), and temperature difference between the long-term mean monthly air and soil (at 40 cm depth) temperatures. The long dashed lines in a, b, and c represent the long-term annual mean values for air temperature, precipitation, and soil temperature at 40 cm depth, respectively. The long dashed line in (d) is the zero value reference.

characteristic is mainly controlled by the presence of the seasonal snow cover during winter months and soil moisture levels caused by relatively heavy rainfall during summer months. The mean monthly air temperatures from November through March are well below 0 °C and precipitation primarily falls as snow. The insulating effect of snow cover prevents the ground from cooling, which elevates the mean monthly ground temperature 12 °C above the mean monthly air temperature. In northern Alaska, the mean monthly ground surface temperature in November can be 15 °C to 20 °C higher than the mean monthly air temperature due to the insulating impact of the seasonal snow cover (Zhang et al., 1997a). Similar results are also reported in European mountain regions (Barry, 1992). Lower soil temperatures in April might result from the impact of snowmelt, while lower soil temperatures during the summer months (May through August) may be due to the effect of rainfall.

Interannual variations of the difference between mean monthly soil temperatures at 40 cm depth and air temperatures were greater (standard deviation of up to ±3.7 °C) during winter, when snow was present on the ground surface, than during summer (standard deviation of about ±1.6 °C) when snow was absent, with the lowest variability occurring in August and September (standard deviation of about ±0.8 °C). This implies that changes in soil temperature during winter did not necessarily respond proportionally to changes in air temperature. The standard deviation of the mean monthly soil temperature at 40 cm depth was greater during winter but its magnitude was reduced compared with the standard deviation of the mean monthly air temperature, while the standard deviation of winter precipitation (snowfall) was relatively small. This suggests that changes in winter air temperature might play an important role in changing winter soil temperatures, but significantly modulated by changes in seasonal snow cover.

5.2. RESPONSE OF SOIL TEMPERATURE TO CLIMATIC CHANGE

The near-surface soil temperature regime is an integrator of all natural processes and their interactions at the ground surface and within the soil. Essentially, all factors affecting the energy and water exchange between the atmosphere and the land surface and within the soil system have an influence on the ground thermal regime.

Changes in the mean annual air temperature (Figure 1) and 40 cm soil temperature (Figure 6a) were of the same magnitude, about 2.0 °C to 2.5 °C, from the beginning of the 20th century to the early 1990s, but the patterns of change were substantially different. The mean annual air temperature increased only slightly and irregularly until the 1960s, while the mean annual soil temperature increased steadily during the entire period.

Figure 9 demonstrates that soil temperatures at 40 cm depth increased between 4 °C and 9 °C during the winter (November through March). There was a decrease of 2 °C to 3 °C in December, January, and February from the beginning of the

1900s to the mid-1920s, then a sharp increase (more than 6 °C) from the mid-1920s to the mid-1930s before a relatively steady rise until the end of the record. In March and November, soil temperatures increased relatively steadily about 4 °C to 5 °C over the entire period of record. During summer (May through August), soil temperatures steadily decreased by 2 °C to 4 °C over the period of record. During the transition months (April, September, and October), soil temperatures showed little or no change. By contrast, air temperature (Figure 10) increased 3 °C to 4 °C during winter (November through March). During summer (May through August), the rate of air temperature change is extremely small with slight increase in May and August and decrease in June and July. During the transition months (April, September, and October), air temperature steadily increased by approximately 2 °C over the entire period. In November and December, air temperature actually decreased 2 °C to 3 °C from the 1900s to the 1950s, and then increased 5 °C to 6 °C from the 1950s to the 1990s.

The soil freezing index at 40 cm depth (Figure 11a) decreased substantially. The decrease (10.3 °C-day/year) of the soil freezing index was more than twice that (4.6 °C-day/year) for the freezing index of air temperature (Figure 2a). The long-term average value of the soil freezing index was about 789 °C-day, just about one-third of the long-term average value of the freezing index of air temperature. The soil thawing index (Figure 11b) decreased over the period of record, while the thawing index of air temperature increased during the same time period (Figure 2b). However, the long-term average values of the thawing indices of air and soil temperatures were still of the same magnitude.

Obviously, both the magnitude and patterns of changes in air and soil temperature are different in this region. There is no one-to-one correspondence in changes of air and soil temperatures either for annual (Figures 1 and 6a) or monthly (Figures 9 and 10) values although the overall magnitude of air and soil temperature increase was about the same during the common period of their records (Figures 1 and 6a). These results demonstrate that changes in air temperature alone cannot account for the changes in soil temperatures in this region.

5.3. RESPONSE OF SOIL TEMPERATURE TO CHANGES IN SUMMER AIR TEMPERATURE AND RAINFALL

Summer soil temperature is mainly controlled by a combined impact of changes in air temperature and soil moisture content. The increase in rainfall during summer months would increase the surface wetness and soil moisture, which results in more energy consumption for evaporation, eventually cooling the ground surface and soils, the so-called soil moisture feedback (Yasunari et al., 1991; Sankar-Rao et al., 1996; Matsuyama and Masuda, 1998). This soil moisture feedback mechanism may explain soil cooling during summer even when air temperatures increased several degrees Celsius over the same period.

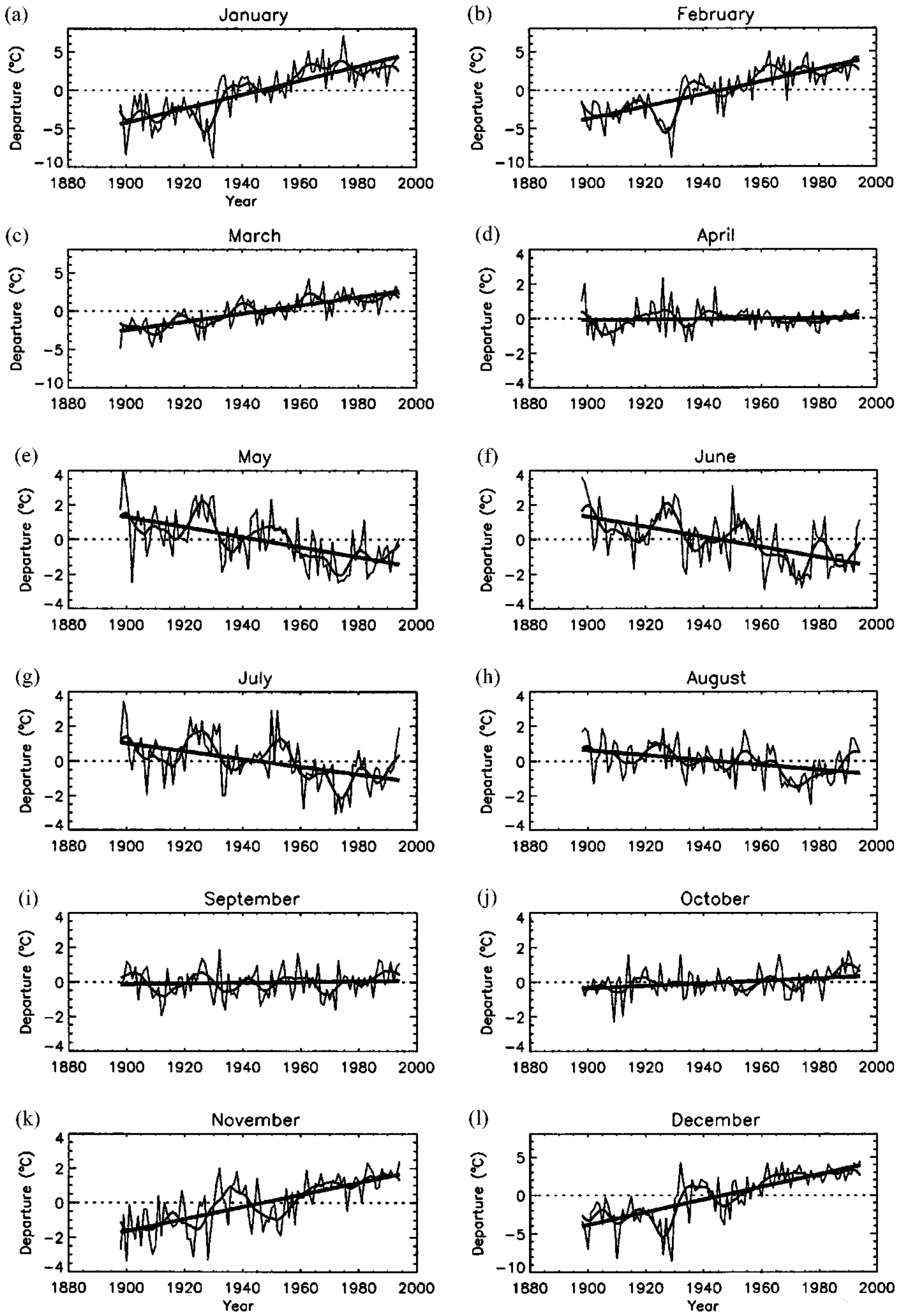


Figure 9. Departure of the mean monthly soil temperatures at 40 cm depth from their long-term mean values for the period from 1898 through 1995 at Irkutsk, Russia. The long-term mean monthly values are shown in Figure 8c. The thin solid line stands for annual values of the departure, while the thick solid line indicates smoothed values of the departure, using a low-pass filter with a cut-off frequency of 0.091. The dashed line is zero value reference.

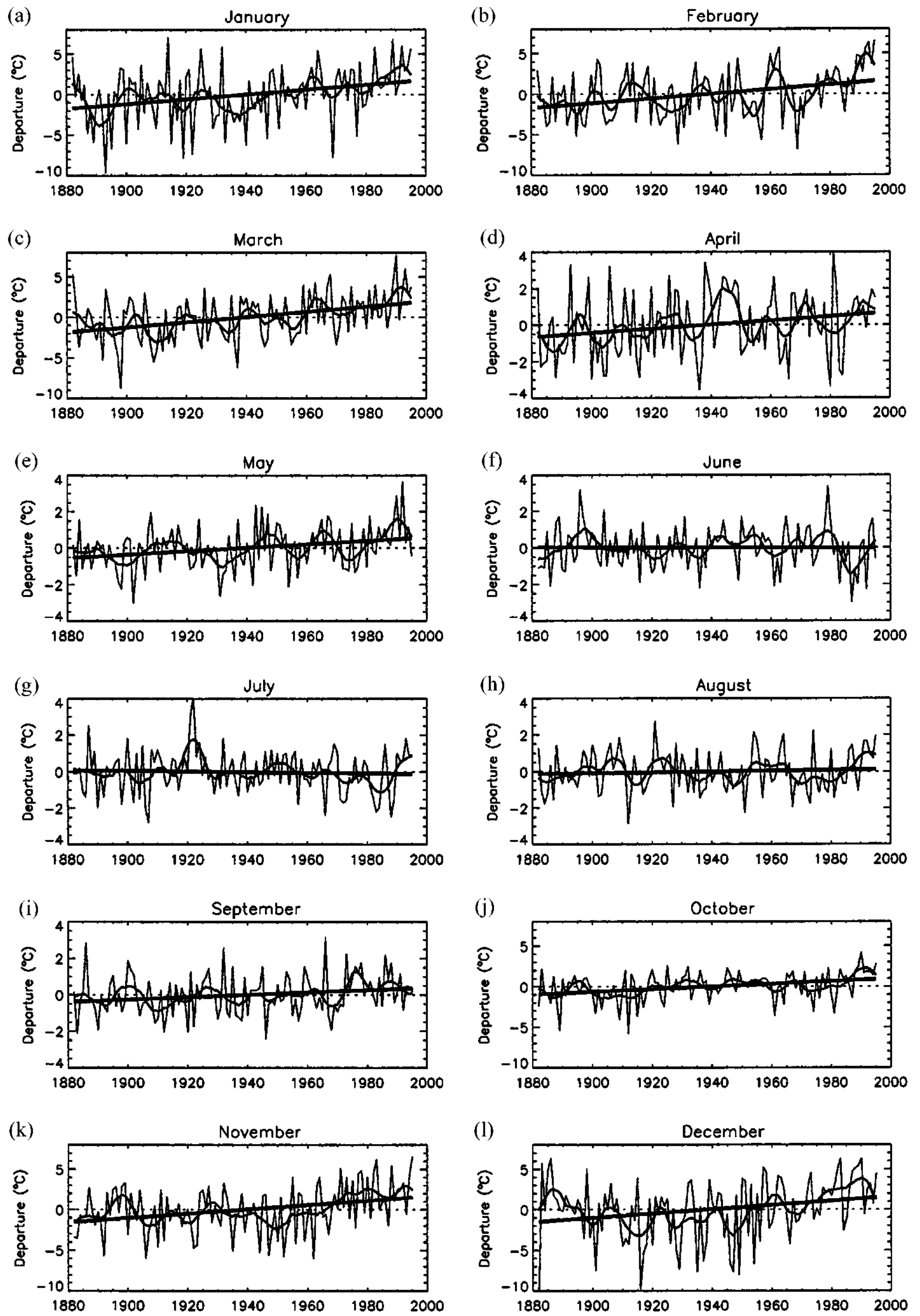
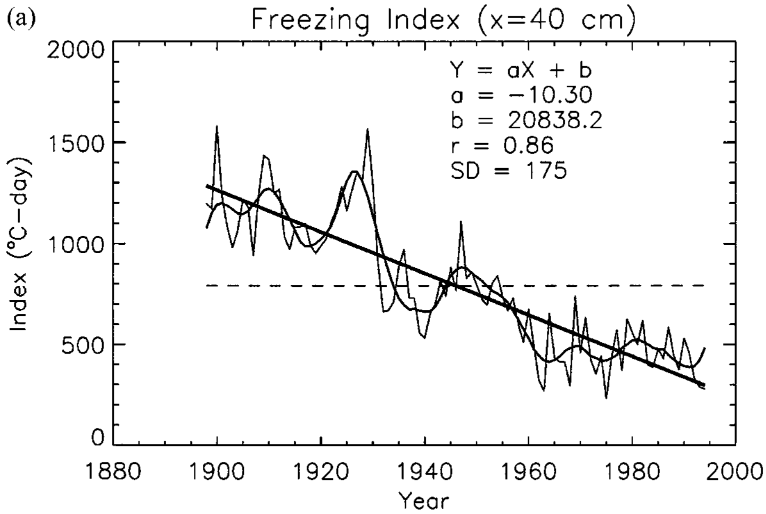


Figure 10. Same as Figure 9 except for air temperature. The long-term mean monthly air temperatures are shown in Figure 8a.

(a)



(b)

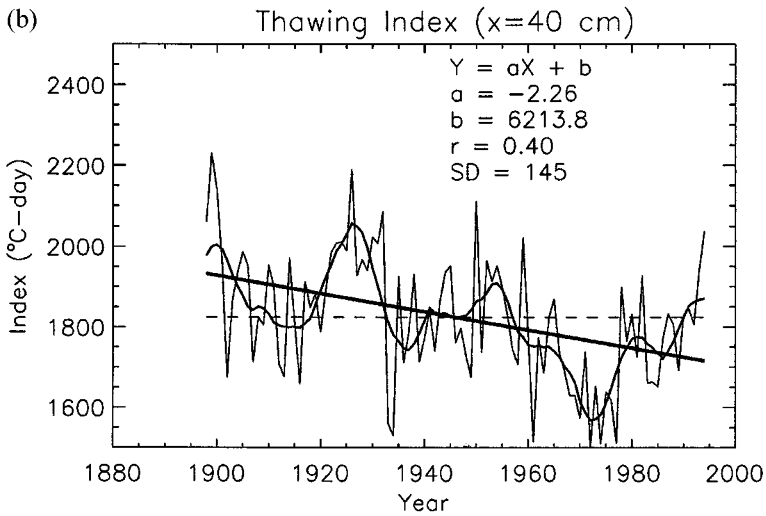


Figure 11. Same as Figure 1 for the freezing (a) and thawing (b) indices of soil temperatures at 40 cm depth. The long-term mean values of the freezing and thawing indices of soil temperatures at 40 cm depth were 790 °C-day and 1824 °C-day, respectively.

There was a clear signal of anti-correlation between precipitation and soil temperature at 40 cm depth, i.e., greater monthly precipitation with lower soil temperature, and *vice versa*. Figure 9g shows that soil temperature in July during the 1920s was well above its long-term average. This relatively high soil temperature is due to the combined impact of relatively higher air temperature (Figure 10g) and lower precipitation (Figure 12g) over the same period. A similar pattern can also be found during the early and mid-1950s. During the 1930s, the July soil temperature at 40 cm depth was below its long-term average (Figure 9g), which may be due to the lower air temperature in July (Figure 10g) and above average precipitation in June and July (Figures 12f and 12g). The lower soil temperature at 40 cm depth during the early and mid-1970s could also have contributed to both relatively lower air temperature and greater precipitation in July. During the late 1970s, air temperature was close to its long-term average in July (Figure 9g), while precipitation decreased substantially from its peak value during the early and mid-1970s in July (Figure 12g). As a result, soil temperature increased sharply over the same period (Figure 9g). These results also imply that precipitation, through the soil moisture feedback mechanism, may play a more important role in influencing soil temperature than air temperature during summer months.

Changes in soil thawing index at 40 cm depth were strongly linked with changes in summer rainfall and air temperature. The soil thawing index decreased from about 2000 °C-day in the early 1900s to about 1750 °C-day from about 1900 to the mid-1910s (Figure 11b). The thawing index of air temperature decreased slightly, from about 2100 °C-day to 2000 °C-day (Figure 2b) which was not enough to account for the decrease in soil thawing index. Over the same period, rainfall increased about 70 mm (Figure 4a). The soil thawing index also decreased substantially from the mid-1920s to the mid-1930s and from the mid-1950s to the early 1970s (Figure 11b), while summer rainfall increased about 200 mm and almost 100 mm over the same periods, respectively, with slight decrease (from the 1920s to the mid-1930s) or essentially no change (from the mid-1950s to early 1970s) in air temperature thawing index. On the other hand, high peaks of soil thawing indices in the mid-1920s and the mid-1950s (Figure 11b) were responding to the low values of summer rainfall with relatively higher thawing index of air temperature. These results again indicate that changes in summer rainfall may play a greater role in controlling summer soil temperature than changes in air temperature.

5.4. RESPONSE OF SOIL TEMPERATURE TO CHANGES IN WINTER AIR TEMPERATURE AND SNOWFALL

The soil temperature increase during winter months appears to represent the combined impact of changes in air temperature and snowfall over the period of record. Air temperature increased much more during winter months than during summer months (Figure 10), which may partly account for the soil temperature increase. However, even if the air temperature increase during winter months were completely transferred into the soil, this accounts for less than half of the soil temperature increase. This implies that changes in precipitation or snow cover during winter months play an important role in soil temperature changes.

The impact of winter precipitation (snowfall) on soil temperatures is mainly through the insulating effect of snow cover. Generally, snowfall decreased about 30 mm from the beginning of the 1900s to the 1930s (Figure 4b). Meanwhile, the freezing index of air temperature increased slightly and was above its long-term average during this period. The combination of less snowfall and colder winters

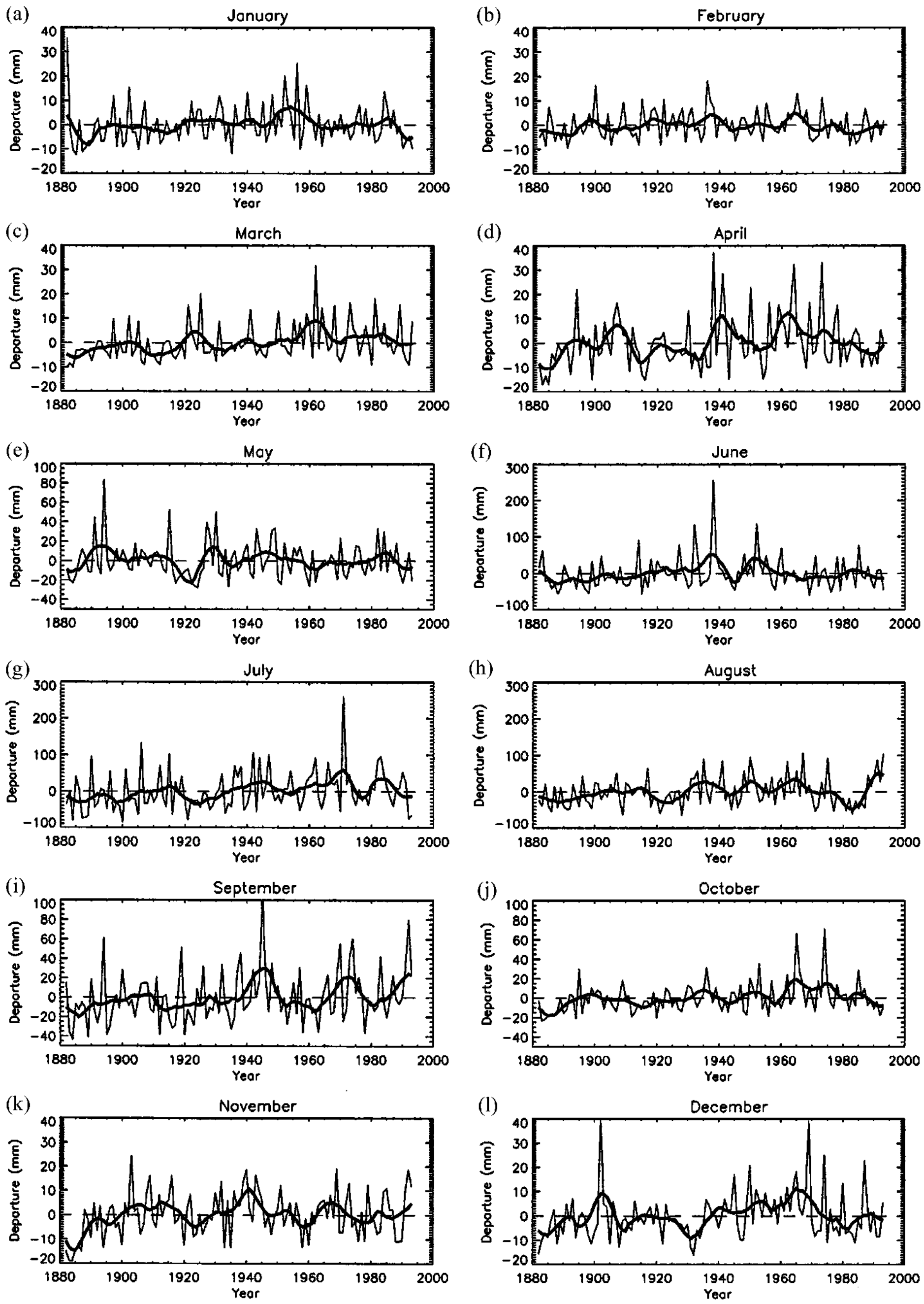


Figure 12. Variations of monthly total precipitation from 1882 through 1993 at Irkutsk, Russia. The long dashed line indicates zero reference; thin solid line represents annual values; thin solid straight line indicates linear regression; and thick solid line shows smoothed values using a low-pass filter with a cut-off frequency of 0.091. The parameters of *a*, *b*, *r*, and SD are the same as in Figure 1 except for precipitation for each month. The long-term mean monthly total precipitation is shown in Figure 8b. Note: y-axis scale for June, July, and August is from –100 to 300 mm; for May, September, and October from –50 to 100 mm; for the other months from –20 to 40 mm.

may account for the soil temperature decrease during winter (December through February, Figure 9) from the beginning of the 1900s to the 1930s. Snowfall increased more than 40 mm from the 1930s to the mid-1960s (Figure 4b), while the freezing index of air temperature was relatively constant except from the late 1950s to the mid-1960s when it decreased sharply and was well below its long-term average value (Figure 2a). Winter soil temperatures increased significantly during this period, especially during the decade of the 1930s (Figure 9a, b, and l). This indicates that during this period changes in snowfall had a greater impact on soil temperatures than air temperature. From the mid-1960s to the end of the record, snowfall decreased by more than 40 mm while soil temperatures during winter months stayed almost constant. The impact of less snowfall, and hence less insulation of the soil may be offset by the substantial increase in winter air temperature (Figure 10) or a decrease in the freezing index (Figure 2a). This suggests that the influence on soil temperature of a change in snowfall of 40 mm is equivalent to the impact of a 600 °C-day change in freezing index at this station, provided that other parameters stayed the same.

The timing of the seasonal snow cover is an important parameter in understanding the influence of snow cover on ground thermal regime (Goodrich, 1982; Zhang et al., 1997a). Early snowfall and snow on the ground provide the most favorable condition for a snow insulating influence on the ground thermal regime (Goodrich, 1982; Zhang, 1993) because early snow cover would prevent the ground surface from cooling for the rest of the winter. Figure 12 illustrates variations of monthly snowfall from 1882 through 1993 at Irkutsk. Overall, increases or decreases of snowfall in October–December (Figures 12a,b) may be the most important factors that affect the soil temperature increase or decrease during the rest of the winter. The abrupt increase in winter soil temperatures in the 1930s might be due to the increase in snowfall at the beginning of winter. The above average snowfall in October through December in the late 1960s and the beginning of the 1970s (Figures 12a–c) may offset the decrease of air temperature when the freezing index of air temperature increased by more than 300 °C-day during the same period. Although there were fluctuations of snowfall in January and February (Figures 12d,e), there was no detectable long-term trend indicating that impact of changes in snowfall on soil temperature during these two months may be negligible.

The duration and thickness of the seasonal snow cover are also important parameters that influence the soil thermal regime (Goodrich, 1982; Zhang, 1993). The reduction in the duration of snow cover over the period of record (1936 through 1985; Figure 5e) was mainly a result of the earlier snow cover disappearance (Figure 5c). Early snowmelt leads to an increase in surface temperature due to the substantial decrease in surface albedo. The snowmelt occurred mainly from mid-March to mid-April. Soil temperature in March increased about 4 °C from the mid-1930s to mid-1980s (Figure 9c), while air temperature increased more than 2 °C during the same period. The limited effect of earlier snowmelt on soil temperature may be offset by the above-average snowfall in March (Figure 12f) since fresh snow has a high albedo and melting snow creates a huge heat sink due to the latent heat of fusion.

Snow thickness is the key determinant of the insulating effect of a snow cover. Field measurements indicate that the impact of changes in snow thickness on ground surface temperature increases exponentially when snow thickness is under 50 cm (Nicholson and Granberg, 1973; Sturm, unpublished field measurements, 1998). When snow is over 50 cm deep, the impact of a change in snow thickness becomes less important. Both average snow thickness (Figure 5b) over the entire winter, and maximum snow thickness (Figure 5d) before the onset of snowmelt in spring, were under 50 cm and decreased from the mid-1930s through the mid-1980s. Decreasing snow thickness should therefore act to decrease winter soil temperatures. Obviously, cooling did not happen mainly because of the early snowfall effect. Another interesting point is that average snow thickness decreased about 10 cm from the mid-1930s to the mid-1980s, while snowfall increased by about 20 mm (water equivalent) from the mid-1930s to the mid-1960s, then decreased more than 30 mm by the mid-1980s. The decreasing snow thickness may also be related to the increasing air temperature during winter through evaporation and internal melting within snowpack. During the day, snow could be melting when air temperatures reach 0 °C or above. Field measurements and numerical modeling results indicate that snow is melting even when the air temperature is a few degrees below the freezing point (Koh and Jordan, 1995; Zhang et al., 1997b).

5.5. EFFECT OF SEASONALLY FROZEN GROUND ON SOIL TEMPERATURE

The influence of seasonal freezing and thawing processes on soil temperatures has received attentions from various investigators (Kudryavtsev et al., 1974; Goodrich, 1978, 1982; Burn and Smith, 1988; Romanovsky and Osterkamp, 1995). Seasonal freezing and thawing influences soil temperatures mainly through differences in the thermal properties of frozen and thawed soils. Changes in the depth and duration of the active layer over permafrost and seasonally frozen soils also have a great impact on soil temperatures. Change in soil temperature due to the impact of seasonally freezing and thawing of soils is often referred as ‘thermal offset’ (e.g., Goodrich, 1982).

Figure 13 shows mean annual soil temperature profiles at Irkutsk in 1929 and 1966. The temperature difference, or thermal offset, between 40 cm and 240 cm depths in 1929 was about –1.4 °C, while the difference in 1966 was essentially zero. The greater magnitude in thermal offset in 1929 might be due to the deeper soil freezing in 1929 than in 1966 (Figure 7d) although the duration of soil freezing in 1929 was shorter than in 1966 (Figure 7c). The primary reason accounting for the big difference in thermal offset between 1929 and 1966 may be the difference in the thermal conductivity ratio between frozen and thawed soils. Using the freezing index and thermal offset values at 40 cm depth for 1929 and 1966, respectively,

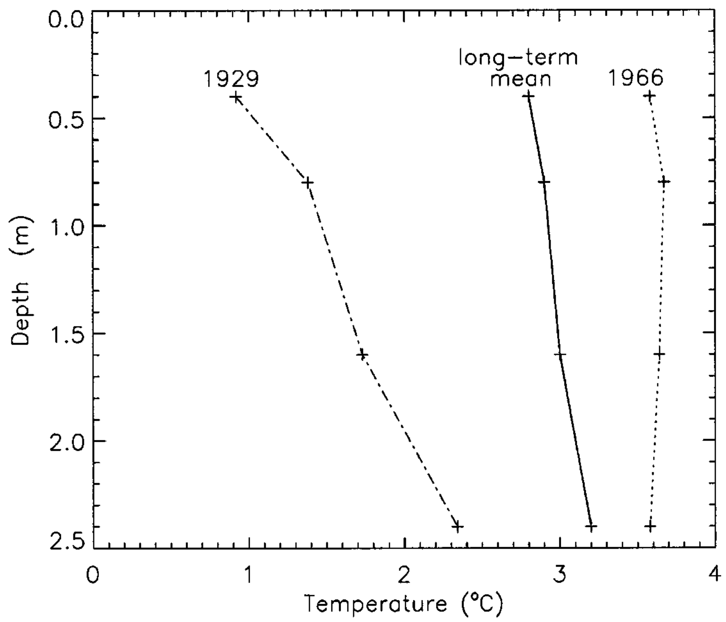


Figure 13. Mean annual soil temperature profiles for 1929 and 1966 and long-term mean soil temperature profile over the period of record at Irkutsk, Russia.

based upon the equation provided by Romanovsky and Osterkamp (1995), the thermal conductivity ratio between frozen and thawed soils in 1929 was about 1.33, while in 1966, the ratio was very close to 1.00. This implies that soil moisture was relatively high before freezing and the thermal conductivity of the frozen soil was about 33% greater than that of a thawed soil in 1929 due to the increase in thermal conductivity from liquid water to ice. In 1966, soil moisture might have been relatively low before freezing but the thermal conductivity of both frozen and thawed soils were essentially the same. Less snow on the ground in 1929 (Figure 4b) may also have increased the thermal offset value due to the reduced soil insulation.

These results indicate that thermal offset can change substantially from year to year depending upon the thermal conductivity ratio between frozen and thawed soils, and the depth and duration of seasonally frozen soils. Changes in thermal conductivity of soils from unfrozen to frozen status may be the most important factor in controlling the thermal offset values. The long-term mean soil temperature profile over the period of record (Figure 13) illustrates that the relatively large difference in long-term mean soil temperatures between 40 cm and 240 cm may be due in part to the existence of seasonally frozen soils at Irkutsk. In other words, soil temperature at shallow depth might be somewhat higher if there were no seasonally frozen soils.

Changes in soil moisture content just before soil freezing in autumn would have significant impact on soil temperatures and soil freezing and thawing. If other conditions are the same, soil freezing (thawing) process may keep soil from cooling (warming) due to the impact of latent heat of fusion. The overall influence of soil freezing and thawing on annual soil temperature may be limited if we assume that the amount of energy used for soil freezing and thawing is the same during the whole seasonal freezing and thawing processes. Interannual variations in soil moisture just before soil freezing may have a substantial impact on soil temperatures. If the other conditions are the same, soil temperature may be lower during years with low soil moisture content just before freezing and higher during years with high soil moisture content just before freezing. However, we do not have soil moisture data at this site and hence further analysis is very limited.

Changes in the first and last day of soil freezing may also have some impact on soil temperatures. The first day of soil freezing at 40cm depth occurs mainly in November (Figure 7a). For the period from 1900 through 1970, soil freezing at 40 cm depth ranged from early November in the 1900s to the middle of November in 1970, while soil temperature in November increased more than 3 °C over the same period (Figure 9k). It is apparent that changes in soil temperature in November directly affect the timing of soil freezing. In general, heat loss from the land surface to the atmosphere is greater, therefore causing soil cooling, for frozen soils than for thawed soils due to the higher thermal conductivity of frozen soils if other conditions are the same. In this case, late soil freezing in 1970s did not enhance the soil cooling in November. The last day of soil freezing mainly occurred in April and early May (Figure 7b). Early thawing of soils above 40 cm depth in the mid-1920s and mid-1940s (Figure 7b) might contribute to the soil warming (Figure 9d) and the late thawing in the mid-1930s might explain the decrease in soil temperature (Figure 9d). Consistent delay of soil thawing at the 40 cm depth from mid-1940s to late 1960s (Figure 7b) may partly explain the 3 °C decrease in soil temperature in May (Figure 9e). There is no clear relationship between soil temperature and duration of soil freezing at 40 cm depth at Irkutsk.

The onset of surface soil thawing starts in April over the grass-covered land surface and complete disappearance of seasonally frozen soils happen in late June or early July at Irkutsk. Existence of a frozen soil layer has a great impact on soil moisture content above the thawed/frozen soil interface. The thin frozen layer would prevent soil water in the thawed layer penetrating deeper into the soil due to its extremely low hydraulic conductivity, hence increasing soil water content. This frozen layer-soil water content interaction would enhance the soil-moisture feedback mechanism and hence cool the soils. Soil temperature in May and June shows large interannual fluctuations and decreased more than in other months from 1898 through 1995 (Figure 9e,f). The frozen layer-soil water content impact may partly explain the relatively strong soil-moisture cooling mechanism. Rainfall is a maximum in July (Figure 8b) with large interannual fluctuations (Figure 12g), while soil temperature decrease was somewhat smaller in July than in May and June (Figure 9e–g) due to the absence of a frozen soil layer in July.

6. Discussion and Conclusion

Long-term soil temperature measurements (1898–1995) have been used to investigate the response of the soil thermal regime to climatic change at Irkutsk, Russia. The primary results can be summarized as follows.

- The long-term mean annual air temperature is about –0.5 °C; values increased by 2.0 °C–2.5 °C from 1882 to the 1990s. Changes in air temperature occurred mainly during the winter months. The air temperature freezing index decreased substantially, corresponding to the significant increase in winter air temperature; the air temperature thawing index also increased slightly.
- The long-term average of annual total precipitation is about 448 mm and annual total precipitation increased by about 130 mm from 1882 through 1995. Changes in instrumentation and observational practice may account for some of this (Groisman et al., 1991). Changes in precipitation occurred mainly during summer months.
- There was a general anti-correlation between the mean annual air temperature and annual total precipitation, i.e., there was more precipitation during cold years and less precipitation during warmer years.
- About 82% of the annual total precipitation falls as rain with the rest falling as snow. Although there were large fluctuations of the rainfall fraction, with extremes ranging from greater than 93% to less than 66%, there were no significant trends of changes in the rainfall fraction.
- There were no significant trends of changes in the first day of snow on the ground in autumn, but snow disappeared progressively earlier in spring, resulting in a reduction of snow cover duration. Both average snow thickness over the entire winter and the maximum snow thickness at the end of winter decreased from the mid-1930s to the mid-1980s.
- The soil at Irkutsk experiences more than nine months of seasonal freezing each year at the surface and about five to six months at 40 cm depth. There were significant interannual variations in the first and last day of soil freezing at surface and at 40 cm depth, which potentially have a significant impact on soil temperatures.
- The long-term mean annual soil temperature is about 2.8 °C at 40 cm depth, 2.9 °C at 80 cm, 3.0 °C at 160 cm, and 3.2 °C at 240 cm. This gradient reflects the impact of seasonal freeze/thaw referred to as the ‘thermal offset’ effect. Patterns of change in soil temperature at all depths were consistent, with the magnitude of change decreasing with depth.
- The magnitude of change in the mean annual air and soil temperature at 40 cm depth were similar over the common period of record (1898–1994), but the patterns of their changes were substantially different. Mean annual air temperature increased slightly until the 1960s, while mean annual soil temperature increased steadily throughout the period. This demonstrates that changes in

air temperature alone cannot explain the changes in soil temperatures in this region.

- Soils cooled by up to 4 °C during summer, while air temperature increased slightly. This cooling in the soil may be explained by changes in summer rainfall. An increase in rainfall would increase the surface wetness and soil moisture, resulting in more energy consumption for evaporation, which leads to cooling of the ground surface and soils. Frozen layer-soil moisture interaction may enhance the soil moisture feedback mechanism, thus cooling soils in early summer.
- Soil temperature increased substantially during winter, by up to 9 °C. This increase may be due to the combined effect of increases in air temperature and snowfall. Air temperature increased about 4 °C to 6 °C during winter. An increase in snowfall during early winter (October and November) and early snowmelt in spring may play a major role in the increase of soil temperatures through the effects of snow insulation and albedo change.
- The long-term average value of the freezing index of soil temperature at 40 cm depth was about 789 °C-day, about one-third of the long-term average value of the freezing index of air temperature. The freezing index of soil temperature at 40 cm depth decreased substantially, with a rate of 10.3 °C-day/year compared with a decrease of only 4.6 °C-day/year for the air temperature freezing index. This demonstrates the substantial insulating effect of seasonal snow cover on soil temperatures.
- The long-term average value of the soil thawing index at 40 cm was about 2090 °C-day and decreased slightly over the period of record, while the thawing index of air temperature increased during the same time period. This indicates that cooling of the soil is due to the increase in rainfall and possibly soil moisture, creating the so-called soil moisture feedback mechanism. However, the long-term average values of the thawing indices of air and soil temperature at 40 cm depth were of the same magnitude, which indicates that the cooling effect of soil moisture is smaller than the insulating effect of the seasonal snow cover on the ground thermal region in this region.
- The overall impact of seasonally frozen soils is to cool the soils in winter due to their relatively high thermal conductivity and in early summer due to the extremely low hydraulic conductivity.

Soil temperature is a valuable parameter with which to monitor climatic change since it integrates all processes occurring at and above the ground surface (air temperature, precipitation, snowfall, seasonal snow cover, vegetation and surface microrelief), as well as the effects of soil type, soil moisture, and freezing and thawing processes. However, changes in soil temperature may not possess a unique signature in response to changes in air temperature and precipitation. For example, increases in air temperature have a positive impact on soil temperature, while the impact of changes in precipitation on soil temperature is more complicated. Increased snowfall in early winter will have a strong positive impact on soil thermal regime, while late snowfall may have a negative impact due to its high albedo and consumption of latent heat during snowmelt. Increased rainfall in summer may also have a negative impact on soil temperatures in seasonally frozen regions due to the soil moisture feedback mechanism. However, in permafrost regions, an increase in rainfall during summer may have little impact on soil temperature, compared to seasonally frozen regions, since the thin active layer over permafrost is almost always at or near saturation due to the extremely low permeability of the underlying permafrost. There may even be a positive impact because the increased moisture content in the peat layer would increase its thermal conductivity significantly.

This study demonstrates that when changes in soil temperature are used as evidence of climatic warming, caution is required because changes in soil temperature are a combined product of changes in air temperature and precipitation, especially snowfall and snow cover on ground. Present findings of the surface warming of permafrost at high latitudes and ground warming at a certain depth below the ground surface elsewhere in the world could be fortuitous and may be misleading since air temperature alone cannot account for such a ground warming. It is safe to state that ground warming is evidence of climatic change (including both changes in air temperature, precipitation, and other climatic variables) rather than climatic warming (mainly indicating increase in air temperature).

This study suggests that global warming may decrease summer soil temperature, especially in the seasonally frozen ground regions and in regions where precipitation falls mainly as rain in summer. Summer soil temperature has a great impact on ecosystem, plant growth, agriculture, heat and water exchange between the land surface and the atmosphere. Therefore, when predicting the potential response of ecosystem to global change, one has to consider the combined impact of changes in air temperature, precipitation, and other climatic variables on soil thermal regime and its effect on the ecosystem. These findings are also important in assessing the results of modeling scenarios of climatic change. They emphasize the need for detailed land surface models incorporating vegetation, snow cover and soil processes.

Further work is planned via a regional analysis using a network of Russian stations where soil temperature records, as well as air temperature, precipitation, snow cover, soil type, and deep ($>$100 m) soil temperature profile data, are available.

Acknowledgements

We wish to express our gratitude to our Russian colleagues who assembled and digitized the soil temperature data. The data release, under the U.S.-Russia Agreement on Cooperation in the Field of Protection of the Environment and Natural Resources, Working Group VIII is scheduled during 2000. We would like to thank Prof. H. N. Pollack and two anonymous reviewers for their constructive comments and suggestions on this article. This research was funded by the U.S. National Science Foundation through the NSF OPP-9614557 and NSF OPP-9907541. Financial support does not constitute an endorsement of the views expressed in this article.

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(Received 17 May 1999; in revised form 3 August 2000)