Temporal and Spatial Variability of Microclimate and Permafrost Conditions in Fairbanks Region, Alaska

By

Galina E. Yershova

RECOMMENDED:

[Signatures]

Advisory committee Chair

Department Head

APPROVED:

[Signatures]

Dean, College of Science, Engineering and Mathematics

Dean of Graduate School

Date
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A

THESIS

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By

Galina E. Yershova, B.S.

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Abstract

Spatial variability of the climate and permafrost conditions in the Fairbanks region was studied using the temperature measurements at a number of sites in the area of approximately 100x100 km around Fairbanks. The effect of climatic and surface parameters (air temperature, vegetation and snow cover, soil properties including water content) on thermal regime of the ground was studied using the temperature and water content measurements at five sites that represent various landscapes and form the Smith Lake profile. The effect of the topographic gradient and slope aspect on the thermal regime of the ground was studied using the temperature measurements and site moisture characterization along three local topo-sequences in the Fairbanks region.

In both considered cases, across various landscapes and along local topo-sequences, the main factor that influences the thermal regime of the ground during the cold period was the snow cover (its depth and duration on the ground) and its combination with the water content on the surface and in the near-surface soils; during the warm period, the main factor was the water content at the surface and in the near-surface soils due to the effect of evaporation from the surface that causes cooling of the ground, and the type of local ground surface vegetation.
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CHAPTER 1

Introduction

Permafrost occupies vast territories in the northern regions. It is linked to the atmosphere by the intervening active layer, vegetation, and snow cover, which vary strongly with time and location. Consequently, it is important to develop a better understanding of the spatial and temporal behavior of the active layer and upper permafrost in response to seasonal, annual, and multi-year changes in climate (Romanovsky & Osterkamp, 1995).

The existence and stability of permafrost is controlled by a set of natural factors, such as climatologic parameters, vegetation and seasonal snow cover, characteristics of the topo-sequences (elevation, topographic gradient and slope aspect) and soil properties (thermal properties and water content). Understanding of the ground thermal regime is extremely important as dynamics of permafrost and active layer can influence the hydrology, hydrogeology and geomorphology of the region and hence the ecosystems and human activities and infrastructure. In light of predicted global warming, this understanding is even more important as warming air temperatures would generate a somewhat different change at the surface of permafrost because of the non-linear effects of the snow cover, vegetation, and active layer (Osterkamp & Lachenbruch, 1990; Nelson et al., 1993). To predict the behavior of the active layer during the projected climatic warming, it is necessary to understand active layer thawing and freezing under natural conditions, which vary significantly from year to year (Osterkamp & Romanovsky, 1997).
Understanding of the ground thermal regime on a local scale (revealing the patterns along local topo-sequences or within different landscapes) allows making more reliable extrapolation of the ground thermal regime over large territories.

The objectives of my thesis research include: 1) revealing the spatial variability of permafrost conditions in the Fairbanks region, 2) revealing the variability of microclimate and permafrost conditions within the topo-sequences: for cases of different elevations and slope aspects, 3) determining the effect of the climate and surface conditions (vegetation cover, snow cover, soil properties including water content) on thermal regime of the ground, 4) determining the main factors that influence permafrost conditions in each case.

Chapter two gives a general description of the Fairbanks area (100x100km area around Fairbanks). It describes the geology, geomorphology, and vegetation and permafrost conditions in the region. Chapter two also gives an overview on climate of the region for the past 100 years, including data on air temperature and precipitation, especially during the winter period.

The third chapter describes the factors that influence the thermal regime of the ground, including vegetation (both upper and lower level), snow cover and its characteristics, soil properties, topographic gradient and slope aspect.

Chapter four presents the analysis of spatial variability of climate and permafrost conditions in the Fairbanks region based on measured data. Detailed analysis on measured temperature data (air, ground surface and ground temperatures) along the local topo-sequences (Nome Creek, Washington Creek and Babe Creek slopes) and general
conclusions on the influence of the topographic gradient and slope aspect on the ground thermal regime is also presented in this chapter.

Chapter five presents the analysis of the effect of climate and surface parameters (air temperature, snow and vegetation cover and soil properties including water content) on the thermal regime of the ground along the Smith Lake profile, consisting of five sites. Detailed description of the profile and analysis of the measured data are included in this chapter.

Chapter six concludes all the results of this research, revealing the main factors that influence the ground thermal regime and climate and permafrost conditions in the Fairbanks region in general, along the local topo-sequences and along a profile within it.

I contributed the major part of the data collection, processing and analyzing and thesis writing.
CHAPTER 2

Environmental Conditions within Fairbanks Area

2.1 Physiography and Geology

Fairbanks is located about 160 km south of the Arctic Circle, on the north side of the broad Tanana River Valley, near the base of the hills of the southern Yukon-Tanana Upland (Figure 2.1). This east-trending upland between the Yukon and Tanana River is an area of rounded hills and ridges 600 to 900 m in elevation. Scattered groups of higher mountains project above ridges to altitudes of 1500 to 1800 m. Summits in the Fairbanks area (approximately 100x100 km around Fairbanks) reach elevations up to 1000 m, but usually not more than 600-700 m, and local relief ranges from 100 to 400 m (Pewe, 1982).

South of the Yukon-Tanana Upland is the wide Tanana River lowland, a sediment filled trough between the uplands to the north and the towering, glaciated Alaska Range to the south. Huge alluvial fans extend northward from the Alaska Range to the vicinity of Fairbanks (Pewe, 1982).

Tanana is the lowest point in the Tanana Valley and has an elevation of about 100 m above sea level. The Chisana River, which is the largest headwater tributary of the Tanana, originates in the glaciers of the Alaska Range. The Tanana River flows in a general north-west direction for an air-line distance of 600 km to join the Yukon, but the actual length of the river is much greater as the course is far from direct. All the large tributaries of the Tanana River enter from the south, and most of the 15-20 rivers that constitute this drainage are mud-laden glacial streams that drain the north flanks of the
Alaska Range. The tributaries from the north, which drain the Yukon-Tanana region, are smaller, clear-water streams.

Break up dates for Tanana River at Nenana range from April 26 to May 17, with an average date of May 7. At Fairbanks the earliest and latest dates of the break-up are April 26 and May 14, with an average date of May 6 (Mertie, 1937). The Tanana begins to freeze about the middle of October, but the small streams in general freeze earlier in the fall and open earlier in the spring than the main rivers.

Pewe (1982) subdivided physiography and geology of the Fairbanks area into loess-covered bedrock hills, lower hillslopes and creek-valley bottoms, organic-silt lowlands at the base of hills and the Chena and Tanana River flood plains. Bedrock hills consist mainly of schist, slate, and gneiss, which are intruded by quartz veins ranging in thickness from less than 2.5 cm to 60 cm. Some hills are composed of granitic rock, and dark basalt crops out locally. Less than 90 cm of loess cover bedrock on hilltops and steep slopes. Loess thickness increases on middle and lower hillslopes. Loess is a homogeneous deposit of silt-size particles from 30 cm to 60 m thick transported by strong winds from the Tanana River flood plain (Pewe, 1982).

Stream-deposited silt fans and small areas of organic silt and peat occur in the valley bottoms. Lower slopes and organic-silt lowlands are perennially frozen and contain large masses of ice. A 30 cm to 4.5 m thick layer of mica-rich, sandy silt covers the surface of the flood plains of the Chena and Tanana Rivers. Sandy gravel of variable thickness lies beneath this silt mantle. Cobbles up to about 7.5 cm in diameter are common. Alternating lenses or beds of silt, sand and gravel extend to a depth of about
200 m. Permafrost is present in the flood-plain sediments, but is generally absent under rivers and lakes. Frozen flood-plain sediments contain ice in the form of intergranular fillings rather than the large ice masses in retransported silt of the lower slopes and lowlands (Pewe, 1982).

According to Capps (1932) and Pewe (1975), the Fairbanks area has not been glaciated, but glaciers from the Alaska Range on the south probably approached to within 80 km. During glacial advances a total of several hundred meters of sand and gravel was deposited in the Tanana Valley by the heavily loaded Tanana River. Aggradation of the Tanana River caused clear-water tributaries from the unglaciated upland to the north to aggrade their lower valleys. More than 90 m of sediment was deposited in creek valleys in the Fairbanks area.

Despite the low rainfall, the Fairbanks area has abundant lakes, swamps and marshes. Except on hilltops, steep slopes, and cultivated land, the ground is wet almost everywhere throughout the summer. A luxuriant spongy mat of low vegetation, mosses with sedges or small shrubs, restricts surface-water movement and acts as a reservoir (Pewe, 1954).

2.2 Climate

Fairbanks has the most rigorous climate of any city in the United States. Weather records from 1904 to 2002 indicate that the region has a continental climate characterized by an extreme range between summer and winter temperatures. The minimum recorded temperature for the Fairbanks area is -54°C and the maximum is 37°C and the mean
annual number of days with freezing temperatures is 233. But freezing temperatures have been reported during every month except for July.

Mean annual air temperature is not the best measure of duration and intensity of cold. A more accurate method uses the freezing index, which is the number of degree days during a freezing season. The degree days for any one day equal the difference between the average daily air temperature and 0°C. The average annual freezing index for Fairbanks based on the air temperature records from 1970 to 1997 is 2940 degree days.

Although summers in central Alaska are the warmest in the state, the mean summer temperature (June, July and August) during the last 30 years is 15.6°C according to data from Fairbanks International Airport. The mean annual air temperature for the same period is -3°C. The mean last day with the freezing temperature in spring is May 22, and August 31 is the mean date for the first autumn frost.

Benson and Rizzo (1979) came to a conclusion that topography generates important gradients in local climate as a result of strong temperature inversions (coldest air temperature in valley bottoms and increasing temperature with elevation) during the calm winter cold periods. In the Fairbanks area, strong inversions occur as much as 80% of the time during cold spells and may produce gradients of up to 21°C/100 m of elevation. These inversions are occasionally broken up by winds from the Alaska Range but can persist for weeks at a time during winter. Temperature inversions are most pronounced during cold high-pressure events in mid-winter, but also occur frequently at night during summer. During the day in summer, inversions are replaced by a normal
adiabatic rate of cooling in which air temperature decreases with increased elevation (somewhat less than 1°C per 100 m) (Slaughter & Viereck, 1986).

According to Slaughter and Long (1974), the large differences in solar input between north- and south-facing slopes result in surprisingly little differences in air temperature. They explain it by convection that effectively mixes air within the canopy with the bulk atmosphere. At the middle of a north-facing slope in CPCRW (Caribou Poker Creek Research Watershed) at 480 m elevation, the mean annual air temperature is only 1°C colder (-2.7 vs. -1.6°C) than at the same elevation of the south-facing slope.

According to Hinzman et al. (in prep.), annual precipitation in interior Alaska is low, from 250 to 500 mm; with a 50-year average for Fairbanks of 287 mm. Precipitation in Fairbanks varies 3-fold among years, from 478 mm in 1990 to 142 mm in 1957. Summer and winter precipitation is generated from major frontal systems that cross the State, but convective storms add significantly to the summer precipitation. Precipitation events in the early summer months (May, June and early July) are typically light and showery with high spatial variability. The relatively dry summer conditions are replaced by the fall rain events, which can be heavy and sustained. On average, precipitation increases throughout the summer. There is considerable variability in annual precipitation in Alaska with low precipitation years, such as 1957, generating frequent wildfires, while high-precipitation years, such as 1967, can result in flooding (Hinzman et al, in prep). Although precipitation during the growing season may be low, evaporation rates are also low because of the relative short growing season and cool temperatures. Even so, as
much as 75 to 100 percent of the summer precipitation may be lost as evapotranspiration (Gieck & Kane, 1986).

Snow acts as an important climate and ecological factor in northern regions. Based upon snow data from Fairbanks International Airport Weather Station, for the last decade, the ground is usually snow-covered from early- mid-October until late April to early May. In some years, the snow period can be considerably longer. For example, in 1992, snow remained on the ground until mid-May and returned permanently on 12 September giving less than a 4-month snow-free period.

Although snow accumulates during the entire winter period, maximum snow pack amounts are generally shallow in interior Alaska (Hinzman et al., in prep). During the last 10 years the average maximum snow pack depth of 67 cm was observed in early-mid April at the Fairbanks International Airport. Several years with heavy snow accumulations took place within the last decade. For example, in 1990-91 and 1992-93 maximum snow pack depth resulted in more than 1 m depth. In contrast there were years with very low snow accumulation, such as 1998-99 and 2000-2001, with maximum snow pack depth of only 46 and 48 cm respectively.

Using precipitation data from the Fairbanks International Airport (1948-2000), Hinzman et al., (in prep.) concluded that snow fall accounts for approximately 35% of the yearly total precipitation. Due to typically below freezing temperature throughout the winter, little snowmelt occurs until spring. During the snowmelt period, (generally late April) snow is released as stream-flow over a relatively short period, making snowmelt the major hydrologic event of the year. Incoming solar radiation is the major factor
governing snow melt as the albedo of the snow decreases from approximately 0.75-0.90 for the fresh snow to 0.40-0.70 for the old snow to 0.10-0.20 for the snow-free vegetation (Hinzman et al., in prep).

2.3 Vegetation

The valleys and lower ridges of the Yukon-Tanana region have a heavy mantle of vegetation, which includes coniferous and deciduous trees, many kinds of flowering plants, ferns, mosses, lichens, and still lower forms of plant life (Mertie, 1937).

Benninghoff (1952), in a study of root habits and plant distribution in the subarctic, concluded that the success of trees growing on soils with permafrost at shallow depth is related to the flexibility of their root habits. Black spruce with its shallow and flexible root system was considered best suited. White spruce, larch, and white birch have shallow but less flexible root systems; and although they have wide ranges on these soils, Benninghoff believed them to be somewhat less frost-tolerant.

According to Mertie (1937), the common trees in the Yukon-Tanana region are the black and white spruce, the balsam poplar or cottonwood, and the quaking aspen. At favored localities, good stands of white birch are also found.

Timber grows in this region up to an average elevation of about 750 m above sea level, but local conditions cause considerable variations in the timber line. Near timber line the spruce becomes smaller and above timber line gives place to low shrub, of which the alder in moist ground is the most common type. Above this are prostrate plants, such as the lichens and certain flowering plants (Mertie, 1937).
Tussocks are characteristic plant form in tundra regions but they also are common in swamps. Tussocks are a common type of vegetation in the Fairbanks area. Hopkins and Sigafoos (1951) described the tussock as a tufted plant form characteristic of certain grasses and sedges growing in areas in which congeliturbation is active. It consists of a ball-like mass of leaving and dead plant parts that stands as a small mound or hummock above the ground surface in areas where the water table and silty mineral soil are close to the surface. The tussock is a single plant; its leaves, culms (jointed stems), and roots give it characteristic form. Small year-old tussocks consist of small tufts without winter-killed leaves. Culm bases rest directly on mineral soil. The older, well-developed tussocks are mounds nearly spherical to vertically elongate in shape and are 5 to 30 cm high. Dead and living culms and leaves cover the surface and give the tussock its ball-like form. The larger tussocks are spaced at intervals of 5-10 to 60 cm and are surrounded by a thick mat of mosses or by low areas in which vegetation of equivalent size is lacking. If thick mats of moss are lacking, the soil surrounding the tussocks is bare or is covered with humus and low mosses forming a layer less than 3 cm thick (Hopkins & Sigafoos, 1951).

The culms, which are the only parts that can increase the height of the plant by growth, form only a small fraction of the height of the tussock. Hopkins and Sigafoos (1951), suggested two hypothesis that might explain this height: the culms started at the level at which they are found and the surrounding soil was removed by erosion, or the culms started at a lower level and were pushed upward to their present position by frost action.
Benninghoff (1952), in his work on interaction of vegetation and soil frost phenomena concluded that vegetation shields the soil from maximum penetration of heat by shading, by decreasing air circulation, by retaining soil moisture in and just above the soil, and by intercepting the rain. Another cooling effect, resulting from evaporation of moisture on plant surfaces, may be significant.

2.4 Permafrost conditions

Most of the Alaskan boreal forest is underlain by discontinuous permafrost. Osterkamp and Romanovsky (1999) concluded that in interior Alaska permafrost temperature varies seasonally within the upper 10 to 15 meters. Permafrost temperatures in the Alaskan boreal forest are 0 to -4°C and typically warmer than -2°C, reflecting the mean annual air temperature of the region (-2 to -5°C).

According to Pewe (1982), permafrost is present nearly everywhere in the Fairbanks area except beneath hilltops and moderate to steep south-facing slopes. The Tanana River flood plain is underlain by perennially frozen ground interspersed and interstratified with zones of unfrozen sediments. Gently sloping alluvial fans and colluvial slopes extend from the upland to the flood plain and creek-valley bottoms and are underlain by continuous permafrost with abundant, large ground-ice masses.

Flood-plain sediments are perennially frozen, often in more than one layer, to depth of up to 80 m (Pewe, 1982). The thickness of these frozen layers varies considerably. Permafrost is absent beneath existing or recently abandoned river channels, sloughs, and lakes, and elsewhere layers of frozen sand and silt are intercalated with
unfrozen layers of gravel. Because gravel layers are commonly lenticular in shape, no single unfrozen layer of broad lateral extent exists.

Depth to permafrost in undisturbed areas ranges from 60 to 90 cm in the older parts of the flood plain to more than 120 cm on the inside of river meanders (Pewe, 1947).

Ice commonly occurs as granules and cement between mineral grains in perennially frozen sediments of the flood plain. The large ice masses common beneath lower hillslopes are not found in flood-plain deposits (Pewe, 1954).

Permafrost in alluvial fans, colluvial slopes and silt lowlands probably extends from the flood plain to the hills and is absent only under larger lakes. The apices of the broad, gently sloping, coalescing alluvial fans extend into the upland valleys, in some places almost reaching the hillcrests. Low-angle silt aprons on lower hillslopes and between alluvial fans contain continuous permafrost, and small lowlands of organic-rich silt and peat that extend from the toe of the fans to the flood plain just north of College Road and in Goldstream Valley are also underlain by continuous permafrost. Permafrost may also occur in isolated, small bodies near the contact with permafrost-free slopes (Pewe, 1982).

Perennially frozen ground in silt fans and beneath slopes is at least 50 m thick near the flood plain, thins toward the hills, and pinches out at the base of steep, south-facing slopes. On north-facing slopes, it may extend to the summits.
Permafrost is encountered 0.6 to 1.2 m below the ground surface on lower slopes and in the creek-valley bottoms and 1.5 to 6 m below the ground surface near the contact with permafrost-free slopes (Pewe, 1982).

Permafrost in fans, slopes, and lowlands contain large horizontal or vertical sheets, wedges and saucer-shaped and irregular masses of ice. Pewe (1982) considered these ice bodies to be up to 4.5 m in thickness and from 0.3 to 15 m in length. Although some ice is clear, some contains silt particles that impart a gray color. The ice is often arranged in a honeycomb network that encloses silt polygons 3 to 12.2 m in diameter, and subsurface polygons produce a polygonal surface pattern in some areas. Ice masses occur at depths of 1.5 to 7.6 m in fans and on colluvial slopes and 30-45 cm to 1.5 m in silt lowlands (Pewe, 1982).

The mean annual ground surface temperatures in the Fairbanks area are usually 3 to 6°C warmer than the mean annual air temperatures. As a result of relatively warm air temperatures and the effect of snow cover in reducing winter heat loss from soil, the mean annual ground surface temperatures in the Alaskan boreal forest often exceed 0°C and can be as high as 4°C. The warm temperature and sensitivity to surface properties make permafrost in the Alaskan boreal forest vulnerable to climate change and disturbance (Osterkamp & Romanovsky, 1999).

Pewe (1982) considered thermokarst features and vegetation to be natural indicators of permafrost. Although thermokarst mounds or pits in cleared fields help define the permafrost boundary, the absence of thermokarst phenomena does not necessarily indicate the absence of permafrost. This is especially true when undisturbed
ice masses lie below the depth of seasonal thawing, or when frozen ground does not contain massive ice.

Pewe (1982) revealed that the boundary between slopes underlain by permafrost and slopes that are permafrost-free is usually marked by a noticeable change in vegetation. A stunted forest of black spruce with knee-high shrub of dwarf birch, Labrador tea, and blueberry, and a thick, moist carpet of cranberry, cottongrass, moss, and caribou lichen grow on poorly drained, gentle slopes underlain by shallow (40 cm to 1.2 m below ground surface) permafrost. Willows and alders follow faint water courses.

White spruce, birch, quaking aspen, and alder are found on better drained, slightly steeper slopes where permafrost is absent or present at a depth of more than 1.2 m. The boundary between black spruce – shrub forest and white spruce - birch - aspen forest is distinct and readily recognizable. Generally, permafrost, with or without ice masses, extends a short distance upslope into the white spruce – birch – aspen forest. The boundary between permafrost and permafrost-free areas is higher on north-facing slopes that receive less solar heat than on sunnier south-facing slopes (Pewe, 1982).
CHAPTER 3

Environmental Factors affecting the Permafrost Temperature Regime

3.1 Vertical temperature profile

A schematic diagram of typical changes of mean annual temperature change with depth for cases of seasonal thawing and freezing of the ground is represented in the Figure 3.1. The mean annual temperatures are starting from the air (2 m above the surface) and going down towards the deeper soil layers.

Within the layer of air adjacent to the Earth’s surface (from 0 to 2 m in height) an increase in mean annual temperatures in the direction towards the ground surface is observed. Various surface covers (lower level vegetation, snow cover) can exert either cooling or warming effect on the ground surface temperature. Their effect will be discussed later in this chapter.

The mean annual temperature in the active layer decreases with depth from $T_{\text{surf}}$ to $T_{\text{mean}}$ (first inflection of the curve in the Figure 3.1). The reason for that is the fact that the thermal properties of the active layer change during the year because of seasonal thawing and freezing. The difference in thermal conductivity between thawed and frozen ground causes a so-called “thermal offset”, which is defined as the difference between the mean annual temperature at the bottom of the active layer and at the ground surface (Kudryavtsev et al., 1974; Goodrich, 1978: Burn & Smith, 1988).

Of importance is the fact that in case of a stable thermal regime, the heat balance exists at any depth of the annual temperature fluctuations, i.e. the amount of incoming heat during the warm half-period equals the amount of outgoing heat during the half-
Figure 3.1 The mean annual temperature change with depth for seasonal thawing (a) and freezing (b) of the ground: $T_{air}^{\text{mean}}$, $T_{surf}^{\text{mean}}$, $T_{\text{mean}}^{f}$ are the mean annual temperatures of the air, at the ground surface and at the base of the seasonal thawing (freezing); $H_{an}$ is the depth of annual temperature fluctuations; $H_{BP}$ is the permafrost thickness (Modified from Yershov, 2001)
period of cooling. Hence, in case of soils with equal thermal conductivity in thawed and frozen states there will be no thermal offset (Yershov, 1998).

The thermal offset is mainly determined by the thermal conductivity. The effect of other thermal properties of soil (heat capacity and latent heat) is practically negligible. According to calculations and experimental results, thermal offset varies in the range of 0-3.5°C (Osterkamp & Romanovsky, 1999).

The thermal offset is usually the largest within a peat layer. Summer temperature conditions on the ground surface also have a significant influence on the thermal offset value (Romanovsky, 1989; Romanovsky & Osterkamp, 1995). Goodrich (1982) concluded that the magnitude of the thermal offset is determined by the active layer thickness. Zhang et. al. (1997) showed that thermal offset increases with increase in the length of the thaw season and maximum thaw depth.

For homogeneous semi-infinite soil system, Romanovsky and Osterkamp (1995) have shown that thermal offset is proportional to thawing index of the ground surface temperature but inversely proportional to the ratio of the thawed to frozen conductivity of soils.

Burn and Smith (1988) suggested that both downward movement of soil moisture in summer and upward movement in winter (Cheng, 1982; Mackay, 1983) may reinforce the asymmetry of seasonal ground temperature profiles.

The ground temperature below the layer of seasonal freezing (thawing) increases with depth due to the geothermal gradient resulting in the second inflection of the temperature curve at the base of the seasonal freezing or thawing layer $T_{\text{mean}}$. If the $T_{\text{mean}} <$
0, the perennially frozen ground is present. One more inflection (the third) of the temperature curve is associated with the lower boundary of permafrost (Figure 3.1a). At this point, unfrozen and frozen material with different thermal physical properties (as a rule \( \lambda_{fr} > \lambda_{unf} \), where \( \lambda \) – is the thermal conductivity of the material in frozen \( [\lambda_{fr}] \) and unfrozen \( [\lambda_{unf}] \) states) is in contact on the lower boundary of the perennially frozen layer. However if this boundary does not move, the thermal fluxes must be equal \( (g_{fr} = g_{unf}) \) in frozen and unfrozen zones, i.e. \( \lambda_{fr} \text{grad } t_{fr} = \lambda_{unf} \text{grad } t_{unf} \). If \( \lambda_{fr} > \lambda_{unf} \), \( \text{grad } t_{unf} > \text{grad } t_{fr} \), i.e. the temperature increases with depth more rapidly in the thawed zone than in the frozen zone. As a result the temperature curve has a point of inflection at the interface. If \( T_{mean} > 0 \), i.e. there is no perennially frozen ground in the profile, the temperature curve is straight below the layer of seasonal freezing (Figure 3.1b) (Yershov, 1998).

3.2 Influence of vegetation on microclimate and thermal regime of the ground

There are several aspects to the influence of vegetation cover (both upper level - trees and shrubs and lower level - grass and moss cover) on the thermal regime of the ground:

1) Both upper and lower level of vegetation influences the thermal regime of the ground by changing the heat exchange between the soil and the atmosphere.

- During the summer, the vegetation cover partially reduces the amount of direct and scattered solar radiation, which leads to the cooling of soils. During the winter, the opposite effect takes place, as vegetation cover works as thermal insulator, reducing the heat flux from soils, which results in soils warming (Kudryavtsev et al., 1974). The total effect of these two processes is dependent on a number of factors and conditions that
determine the role of vegetation cover in heat exchange between the soil surface and
atmosphere and between soil surface and underlying soil and rocks, such as duration of
summer and winter seasons, continentality of climate, snow cover depth, moisture
content in the underlying soil etc. (Kudryavtsev et al., 1974).

2) Upper and lower level vegetation influences the thermal regime of the
ground by changing the moisture exchange between the atmosphere and soils in the
following ways:

✓ Evaporation and evapotranspiration from soils varies for different types of
vegetation. This changes the moisture content in the air and in this way influences the
thermal balance between the air and soils and the thermal properties of soils (Yershov,
2001).

✓ Vegetation, especially tree growth, intercepts a significant amount of
atmospheric precipitation, both rain and snow, by as much as 10 to 40%. Any rain carries
some amount of heat, so interception of rain results in reduction of heat entering the
ground. Interception of snow decreases snow cover thickness under the trees. At the same
time interception of precipitation results in decrease in water content in soils (Brown,
1963).

✓ The ability of vegetation to retain the moisture in soils is strongly
dependent on the type of vegetation and in this way it modifies the thermal properties of
soils (Kudryavtsev et al., 1974).

✓ Vegetation cover (especially trees, shrubs and tussocks) determines
conditions of snow accumulation, redistribution, properties of the snow layer and the
duration of a period when snow is present on the ground. Once again, the magnitude of these effects is dependent on the vegetation characteristics (Kudryavtsev et al., 1974).

3) According to Tyrtikov (1979), in natural conditions, not only does the vegetation influence the thermal regime of the ground, but also the reverse effect takes place. So permafrost and vegetation cover evolve together, influencing one another.

Yershov (1998) concluded that in the area of perennally frozen ground the influence of vegetation cover on the depth of seasonal thawing is greater than that on the depth of freezing in the area where permafrost is absent. In both cases removal of vegetation cover leads to higher amplitude of annual temperature fluctuations at the ground surface with both lower minimum and higher maximum temperatures. According to Kudryavtsev et al., (1974), the presence of grass-moss cover, shrubs and forest decrease the mean annual amplitudes of temperature in soils by 15-25% taken separately and by 30-50% working together.

In the continuous permafrost zone, where cold permafrost prevails, the effects of vegetation are reflected largely in differences in thickness of the active layer and do not appear to change significantly the thermal regime of permafrost. But in the discontinuous permafrost zone, where ground surface temperature is close to 0°C, vegetation may influence the thickness of the active layer and may determine whether permafrost is present or absent (Luthin & Guymon, 1974).

Benninghoff (1966) points out that where the surface mean annual temperature is not colder than -2°C or -3°C, the kind and condition of the vegetation cover will be critical to the development and stability of permafrost.
3.2.1 Effect of different vegetation types

The larger the phytomass surface of the forest, which depends on height, density and closeness of its layers, the less the solar rays penetrate to the soil surface. According to Tyrtikov (1959), radiation in an open area during summer can be two orders of magnitude higher than that under a dense forest canopy. In winter, the forest hinders the rate of radiation emitter from the ground but the effect is not as pronounced because of reduced foliage.

Forest and shrub vegetation reduces the inflow of the solar energy to the surface due to the effect of shadowing. As a result, the surface warms less in summer and the snow melting period becomes longer compared to the forest-free sites (Yershov, 1998).

Shrubs play an important role in snow distribution and redistribution. McFadden et al., (2001) in the work concerning interaction of shrubs and snow in the arctic tundra mentioned that snow depths correlate closely with shrub canopy height and stem diameter. According to their observations, shrubs increased snow depths by 27%, independent of local variations in topographic relief. McFadden et al., (2001) suggested that an increase in shrub cover could significantly increase snow depths in the region, even without an increase in precipitation, as shrubs increased snow accumulation by an amount approximately equal to the fraction of the total winter snowfall that is normally lost to sublimation.

Grass cover causes to a lesser degree changes in the heat exchange between soil surface and the atmosphere and the ground temperature regime. According to Yershov (1998), the total thermal effect of grassy vegetation on mean annual ground temperature
can be either warming or cooling, but it does not exceed fractions of a degree C. The seasonal amplitude of mean monthly temperatures also reduces insignificantly.

The most important influence on thermal regime of the ground is produced by ground vegetation covers, such as moss or moss-lichen, that act as natural thermal insulators, preventing heating of the soil in summer and reducing heat yield from the surface in winter. This can be explained by a considerable change in thermal conductivity in humid, natural soil covers while passing from the thawed state into the frozen (Gavriliev, 1998; Beringer et al., 2001). According to Yershov (1998), thermal conductivity of moss—lichen cover in the thawed state is 0.1-0.7 Wm\(^{-1}\)K\(^{-1}\), one third to one half of that in frozen state. Feldman et al. (1988) suggested that in winter time the thermal conductivity of moss can be as high as 1.8 Wm\(^{-1}\)K\(^{-1}\), due to the fact that moss has a large capacity for retaining moisture. If the water content in moss is high at the moment of freezing, then its thermal conductivity will increase significantly in the frozen state and hence moss will allow the cold waves to penetrate deep into the ground. In case of low water content in the moss before freezing, values of the thermal conductivity in thawed and frozen state will be very similar.

During summer, when evaporation rates are high, moss loses moisture and its thermal conductivity can decrease to 0.2 Wm\(^{-1}\)K\(^{-1}\) and possibly even lower (Beringer et al., 2001). For this reason, moss is an effective insulator during summer. Consequently, permafrost is more stable and the active layer is shallower in areas with moss.

According to Yershov (1998) the ability of moss cover to retard the entry of heat in summer period is greater than its ability to restrict the yield of heat in winter by the
same order. Thus, the layer of moss 2-3 cm thick reduces the sum of thawing degree-days by two-thirds and more.

3.3 Snow cover

3.3.1 Characteristics of snow cover

Characteristics of snow cover (depth, density, structure and texture) are continuously changing in time. These changes take place due to some internal factors (difference between surface and atmospheric conditions that would not be taken in consideration in this work) and external factors: consolidation (due to its own weight), wind, solar radiation and liquid precipitation (Pavlov, 1979).

Of more importance is the fact that the density of snow increases mainly because of consolidation due to its weight. The most dramatic changes occur during the initial period of snow being on the ground. For example, in Yakutia, where winters are typically windless and no thaw is usually observed during the winter, the density of fresh snow is not more than 0.05-0.07 g/cm³ and at the end of winter snow density can become as high as 0.2-0.25 g/cm³ (Pavlov, 1979; McFadden et al., 2001).

With the increase in snow depth, the amount of solar radiation reflected from the surface increases, due to the increase in the mass of snow that takes part in scattering and reflection. But one should keep in mind that the maximum scattering takes place in the upper 4-6 cm. Other factors that influence the reflective ability of snow are the structure and water content in snow cover (Pavlov, 1979).

The thermal regime of snow cover is affected by internal factors (its characteristics), that influence the heat transport within it and external – air and ground
surface temperatures, heat and moisture exchange on the boundaries of snow with soil surface and with the atmosphere (Pavlov, 1979; Sturm et al., 2001).

According to Pavlov (1965), during some parts of the day, for example during clear nights, the snow surface temperature can significantly differ from that of the air. According to his measurements in mean monthly terms the difference between the snow surface and air temperature does not exceed tenths of degree with snow surface temperature being colder. On average, during the winter period, the air temperature can exceed the snow surface temperature by not more than 1°C.

Snow depth and its distribution and redistribution during the cold period of time are strongly dependant on the types and density of vegetation cover.

3.3.2 Influence of snow cover on thermal regime of the ground

Snow cover changes the heat exchange between ground and atmosphere considerably. There are several aspects to snow cover influence on thermal regime of the ground.

✓ The albedo of snow is much higher than that of bare soils or vegetation. These results in reduction of solar energy absorption and lower snow surface temperatures compared to that of the soil or vegetation covers.

✓ Due to its lower thermal conductivity (0.05-0.3 Wm⁻¹K⁻¹, which is approximately 5-10 times lower than that of mineral soil) snow cover works as a thermal insulator, preventing the loss of heat by the ground during the cold period.

✓ If after the air temperatures become positive and the snow cover still remains on the surface, snow cover begins to exert the cooling effect on the ground.
surface. In this case snow cover prevents ground surface from heating partially due to reflection of incoming solar energy and partially due to its losses for snow melting. Melting snow maintains the temperature of 0°C at the ground surface despite the positive air temperatures at this time. This leads to a certain cooling of the ground and lowering of its mean annual temperature (Kudryavtsev et al., 1974).

- The total effect of the snow cover results not only in increase of mean annual ground surface temperatures, but also in reducing the ground surface temperature seasonal amplitudes, as compared to that of the air (Yershov, 1998).

- In most cases snow cover may cause the changes in soil water content (Yershov, 1998).

- Kudryavtsev (1978) found out that the greater the heat turnover through the soil surface (the amount of heat going through the soil surface for the half-periods of heating or cooling), the more significant is the snow cover influence on the mean annual temperature and amplitude of surface temperatures (with other conditions being equal). Hence, the most pronounced effect of snow cover occurs at mean annual ground temperatures close to 0°C in conditions of continental climate and very wet soils.

So, the variation in magnitude of the insulating effect of the seasonal snow cover is due to the changes in timing and duration, accumulation and melting processes, thickness, density and structure of the snow cover (Goodrich, 1982; Zhang, 1995; Zhang & Stamnes, 1996), and the interaction of wind, ground surface morphology, soil properties and soil water content and vegetation with the seasonal snow cover (Sturm et al., 2001).
3.4 Soil and mineral properties

The thermal regime of the ground is strongly dependant on soil and mineral properties such as soil composition (including content of salt and organic matter), grain size, structure and texture of soils, density, pore volumes, moisture content (in free and bound states), and hence thermal properties (thermal conductivity, heat capacity and thermal diffusivity).

- Thermal conductivity (with all other conditions being equal) decreases from disintegrated rock to sand, to clay (loamy) sand, to loam, to clay and to peat. Roughly, the depth of seasonal freezing (thawing) is directly proportional to the square root of thermal conductivity. Hence, with all other conditions being equal, the maximum depths of seasonal freezing (thawing) form in rock and coarse-grained material and the minimal in fine-grained soils and peat (Kudryavtsev, 1978);

- Thermal conductivity increases with increase in soil density due to formation of larger amount of heat-conducting contacts in soil.

- Increase in water content in soils leads to considerable increase in thermal conductivity due to the replacement of low heat-conductive air (0.023 Wm\(^{-1}\)K\(^{-1}\)) to higher heat-conductive water (0.57 Wm\(^{-1}\)K\(^{-1}\)) or ice (2.29 Wm\(^{-1}\)K\(^{-1}\)). Hence, thermal conductivity of soil in the frozen state is significantly higher than that in thawed soils.

- According to Kudryavtsev (1978), the depth of seasonal freezing/thawing decreases in general with increase in water content in soils as more heat is consumed/released on phase transition of water;
Soil composition and water content affect the mean annual temperature at the bottom of the active layer due to the effect of the thermal offset. With increase in water content in soils, the difference between the thermal conductivities in the thawed and frozen states significantly increases. This causes the increase in thermal offset and, hence, the decrease in the mean annual temperature on the bottom of the freezing (thawed) layer (Kudryavtsev, 1978). The effects of unfrozen water on the ground thermal regime are the largest immediately after freeze-up and during the period following cooling of the active layer and near-surface permafrost. Effects are smaller during warming and thawing of the active layer. The effects are evident for a few weeks in cold permafrost and lasted for most of the winter period in warm permafrost (Romanovsky & Osterkamp, 2000);

Unfrozen water in the freezing and frozen active layer and near-surface permafrost retards temperature changes because it changes thermal properties and introduces a distributed latent heat effect, protecting the ground from rapid temperature changes (Burn, 1992; 1998a; 1998b). This creates stronger thermal gradients at the ground surface after freeze-up, which increase the heat flux out of the ground and enhance the insulating effect of the snow cover during this period (Romanovsky & Osterkamp, 2000);

According to the results of numerical modeling (Romanovsky & Osterkamp, 2000), the inclusions of the unfrozen water in the freezing active layer and near-surface permafrost with all other conditions being the same, results in ground
surface temperatures up to 9°K warmer (Deadhorse, Alaska site) during the month of freeze-up and the ensuing period of active layer cooling;

✓ Gavriliev (1998) considers the presence of peat in soils to be the most important factor influencing the ground thermal regime and the depth of seasonal freezing and thawing. Physical properties of peat significantly differ from that of mineral soil mainly due to the higher hygroscopic ability of peat. Volumetric water content in peat can reach up to 70-90% compared to 30-40% in the mineral soils. As a result of such high volumetric water content, significant differences in thermal conductivities of peat in the thawed and frozen state are observed. According to Gavriliev (1998), the thermal conductivity of peat in the thawed state varies from 0.29 to 0.52 W/(m°K) and in the frozen state can reach up to 1.3 and more W/(m°K). With increase in moisture content in peat, the difference in thermal conductivities of the frozen and thawed peat increases and the ability to prevent the underlying ground from heating in summer becomes greater than the ability to prevent the yield of heat in winter, and the cooling effect increases (Yershov, 1998).
3.5 The effect of topographic gradient and slope aspect on microclimate and permafrost conditions

The position of the site in the topography (absolute elevation and the location in the topo-sequence) and slope aspect to a great extent defines the air, ground surface and ground thermal regime and the depth of seasonal freezing or thawing.

✓ During the warm period, air temperatures decreases with increase in elevation. During the cold period, air temperature inversions take place along the slopes;

✓ Increase in elevation causes changes in soil properties, vegetation and snow cover characteristics (Yershov, 1989);

✓ According to Kudryavtsev (1978), mean annual ground temperature decreases from south-facing slopes through south-western and north-eastern to northern ones. He explains it by the fact that during the cold period the amount of incoming solar radiation is small and equal for slopes of all aspects, but during the warm period warming is stronger on south-facing slopes. This effect is less pronounced in high latitudes;

✓ The influence of slope aspect is highly pronounced on the depth of seasonal thawing and has almost no influence on the depth of seasonal freezing (as it is formed mainly in the winter period when the insolation is practically the same for northern and southern slopes) (Yershov, 1998);

One should keep in mind that the effect of slope aspect works not only due to direct influence of solar radiation and its amount but also by the way of other factors. For example, depth and characteristics of snow cover and its redistribution (depending on the prevailing winds in the area), type and density of vegetation cover at all levels and soil
properties (including moisture content in soils) may vary between north and south-facing slopes due to the influence of solar radiation, and these factors in turn affect the permafrost conditions.
CHAPTER 4

Changes in Air and Ground Temperature Regime along the Local Topo-Sequences

4.1 Spatial variability of climatic and permafrost conditions in the Fairbanks region

Spatial variability of the air temperature in terms of mean annual, mean July and mean January values in the Fairbanks region is represented in the Figure 4.1.

Mean annual air temperatures (MAAT) for the hydrological year 2001-2002 in the region vary from -4.5 to 0.1°C. Differences in MAAT are primarily due to differences in winter air temperatures, that are the most pronounced in January. The only pronounced pattern in the distribution of the air temperatures in the region is along the local topo-sequences. For example, 1.7°C difference in MAAT between the summit and the valley bottom at the Nome Creek slope and 3.2°C at the Washington Creek slope was observed.

Spatial variability of the ground surface temperature in terms of mean annual, mean July (the warmest) and mean January (the coldest) values in the Fairbanks region is represented in the Figure 4.2.

Mean annual ground surface temperatures (MAGST) in the region vary from -3.6 up to +4.3°C. Significant differences in the MAGST are due to significant differences in the winter ground surface temperatures that are the most pronounced in January. Once again the only pronounced pattern in the distribution of the ground surface temperatures in the region is along the local topo-sequences. For example, the difference of 0.5°C in the MAGST was observed between the summit and the valley bottom at the Washington Creek slope, 0.7°C at the Nome Creek slope and 0.8°C at the Babe Creek slope.
Figure 4.1 Variability of the air temperature and active layer depths in the Fairbanks area (2001-2002).
Figure 4.2  Variability of the ground surface temperature in the Fairbanks area (2001-2002)
Spatial variability of the ground temperatures in terms of mean annual, mean March (the coldest) and mean October (the warmest) values in the Fairbanks region is represented in the Figure 4.3. The depth of the ground temperature measurements varied from 0.3 to 0.5 m for different sites within the Fairbanks region.

Mean annual ground temperatures (MAGT) in the region vary from −2.7 to 1.7°. Once again, the only pronounced patterns in the distribution of the ground temperatures in the region are observed along the local topo-sequences. For example, 3.1°C, 2.1°C and 1°C differences in the MAGT, with the colder ground temperatures at the valley bottoms, were observed along the Washington Creek, the Babe Creek and the Nome Creek slopes respectively.

Significant variability of the active layer depths is observed in the region. It varies from 0.3 up to more than 0.9 m. The strongest variability is observed along the local topo-sequences. For example, at the Washington Creek slope, the active layer varies from 0.39 m at the valley bottom to more than 0.95 m at the summit.

In general within the Fairbanks area, the main variability in air, ground surface and ground temperatures is pronounced along the local topo-sequences and there is practically no patterns over the entire region from south to north. However, this conclusion could change when more spatially distributed data become available.

4.2 Influence of topographic gradient and slope aspect on the microclimate and permafrost conditions within the topo-sequences

In order to reveal the variability of microclimate and permafrost conditions within the topographic sequences for cases of different elevation and slope aspect profiles along
Figure 4.3 Variability of the ground temperature in the Fairbanks region (2001-2002)
three slopes: Washington Creek slope, Babe Creek slope and Nome Creek slope were chosen (Figures 4.4 and 4.5).

Slope at **Washington Creek** is of S-SE aspect with absolute elevation varying from 260 to 450 m above sea level. Profile along Washington Creek slope consists of five sites, going from the summit to the valley bottom.

Slope at **Babe Creek** is of N-NW aspect with absolute elevations very similar to those at Washington Creek slope (from 300 to 490 m above sea level). Profile along the Babe Creek slope consists of five sites, going from the summit to the valley bottom.

Slope at **Nome Creek** is of southern aspect with absolute elevations considerably higher than that at Washington Creek and Babe Creek slopes, varying from 690 to 920 m above sea level. Profile at Nome Creek consists of four sites going from the summit to the valley bottom.

Data on topographic position, vegetation, site moisture, surficial geology and soil texture of the sites is shown in the Table 4.1. This data was provided by Teresa Nettleton, a PhD student at the Institute of Arctic Biology and biology department.

Measurements of air temperature were taken on the summits and valley bottoms of slopes at Nome Creek and Washington Creek with HOBO sensors.

Ground surface temperature (at the depth of 0.02 m beneath the moss surface) and ground temperature at the depth of approximately 0.5 m were measured at each site along all three slopes with Optic Stowaway sensors.
Figure 4.4  Schematic display of sites along the Washington Creek and Babe Creek slopes and corresponding slope aspect

Figure 4.5  Schematic display of the sites and slope aspect along the Nome Creek slope
Table 4.1 Description of the sites

<table>
<thead>
<tr>
<th>Site #</th>
<th>Geogr. area</th>
<th>Topogr. position</th>
<th>Vegetation description</th>
<th>Site moisture</th>
<th>Surficial geology</th>
<th>Soil texture</th>
</tr>
</thead>
<tbody>
<tr>
<td>12</td>
<td>65°10.033'N 147°53.648'W</td>
<td>Washington Creek Summit</td>
<td>Semi-open moss and tree dominated woodland</td>
<td>Subxeric</td>
<td>Loess</td>
<td>Silty loam</td>
</tr>
<tr>
<td>13</td>
<td>65°09.984'N 147°53.638'W</td>
<td>Washington Creek Shoulder</td>
<td>Dense lichen/moss dominated woodland</td>
<td>Subxeric</td>
<td>Loess</td>
<td>Silty loam</td>
</tr>
<tr>
<td>14</td>
<td>65°09.928'N 147°53.544'W</td>
<td>Washington Creek Sideslope</td>
<td>Fairly dense, moss dominated woodland</td>
<td>Subxeric</td>
<td>Loess</td>
<td>Silty loam</td>
</tr>
<tr>
<td>15</td>
<td>65°09.478'N 147°52.928'W</td>
<td>Washington Creek Toeslope</td>
<td>Fairly open, moss dominated bog</td>
<td>Mesic</td>
<td>Loess</td>
<td>Silty loam</td>
</tr>
<tr>
<td>16</td>
<td>65°09.271'N 147°51.790'W</td>
<td>Washington Creek Valley bottom</td>
<td>Fairly open, sphagnum dominated bog</td>
<td>Mesic to subhygric</td>
<td>Loess</td>
<td>Silty clay loam</td>
</tr>
<tr>
<td>17</td>
<td>64°59.792'N 147°39.183'W</td>
<td>Babe Creek Summit</td>
<td>Partially open moss dominated woodland</td>
<td>Mesic</td>
<td>Loess</td>
<td>Silty loam</td>
</tr>
<tr>
<td>18</td>
<td>64°59.828'N 147°39.303'W</td>
<td>Babe Creek Shoulder</td>
<td>Fairly open moss dominated woodland</td>
<td>Subxeric to mesic</td>
<td>Loess</td>
<td>Silty loam</td>
</tr>
<tr>
<td>19</td>
<td>64°59.911'N 147°39.404'W</td>
<td>Babe Creek Sideslope</td>
<td>Fairly open shrub and moss dominated woodland</td>
<td>Subxeric</td>
<td>Loess</td>
<td>-</td>
</tr>
<tr>
<td>20</td>
<td>64°59.990'N 147°39.965'W</td>
<td>Babe Creek Toeslope</td>
<td>Partially open moss dominated woodland</td>
<td>Subxeric</td>
<td>Loess</td>
<td>Sandy clay loam</td>
</tr>
<tr>
<td>21</td>
<td>64°59.975'N 147°40.117'W</td>
<td>Babe Creek Valley bottom</td>
<td>Very open moss and shrub dominated bog</td>
<td>Subhygric</td>
<td>Loess</td>
<td>-</td>
</tr>
<tr>
<td>104p</td>
<td>65°21.743'N 146°40.000'W</td>
<td>Nome Creek Summit</td>
<td>Very open shrub dominated tree line</td>
<td>Mesic to subhygric</td>
<td>Loess</td>
<td>Silty loam</td>
</tr>
<tr>
<td>104</td>
<td>65°21.600'N 146°40.418'W</td>
<td>Nome Creek Shoulder</td>
<td>Open shrub dominated woodland</td>
<td>Mesic to subhygric</td>
<td>Loess</td>
<td>Silty clay loam</td>
</tr>
<tr>
<td>105</td>
<td>65°40.243'N 146°40.243'W</td>
<td>Nome Creek Sideslope</td>
<td>Very open shrub dominated woodland</td>
<td>Mesic to subhygric</td>
<td>Loess</td>
<td>Silty clay loam</td>
</tr>
<tr>
<td>104c</td>
<td>65°20.954'N 146°40.123'W</td>
<td>Nome Creek Valley bottom</td>
<td>Fairly open seedling dominated lowland</td>
<td>Mesic</td>
<td>Loess</td>
<td>Loamy silt</td>
</tr>
</tbody>
</table>
4.3 Effect of topographic gradient on air temperature

Air temperature was measured at two slopes with southern aspect. Air temperature at the valley bottoms and summits at Washington Creek and Nome Creek slopes is shown in the Figures 4.6 and 4.7 respectively.

Strong air temperature inversions were observed along both slopes during the cold period. In October, the air temperatures at the valley bottoms and summits of the slopes were almost equal for both slopes. Starting from November till March, mean monthly air temperature in the valley bottom was significantly colder than that on the summits. The maximum difference in air temperatures between the valley bottom and summit at Washington Creek slope was 8°C in December, 2001 and at Nome Creek slope 6°C in November, 2001. Then in April, the air temperatures started to merge and stayed nearly equal through all the warm period.

As we can see from the Table 4.2, considerable differences were observed in both mean winter air temperatures (MWAT) and mean annual air temperatures between valley bottoms and summits along both slopes. Here and after, winter period is from December through February and summer period from June through August. The differences in MAAT between the summits and valley bottoms were 3.2 and 1.8°C for Washington Creek slope and Nome Creek slope respectively. Considerable differences in MAAT along both slopes were formed due to differences in MWAT that were 6.5 for Washington Creek slope and 4.2°C for Nome Creek slope.
Figure 4.6  Mean monthly air temperature along Washington Creek slope, 2001-2002

Figure 4.7  Mean monthly air temperature along Nome Creek slope, 2001-2002
4.4 Effect of topographic gradient on the ground surface temperature

At Washington Creek slope, site moisture and hence the water content in near-surface soils increases from summit to valley bottom from subxeric to mesic-subhygric (Table 4.1). Vegetation down the slope changes from moss and tree dominated woodlands at the summit to moss dominated bogs in the valley bottom.

Measurements of the snow depths were made in the beginning of March, 2002 along the Washington Creek slope. Variability of snow depths were observed within each site but, in general, increase in snow depths occurred in the direction from summit to the valley bottom: from 0.25-0.49 m on summit to 0.4-0.64 m in the toeslope.

Ground surface temperature was measures at all five sites: from the summit to the valley bottom (#12-#15). Variability of the ground surface temperatures is shown in the Figure 4.8. During the period from October to February, mean monthly ground surface temperature was the warmest at the valley bottom and toeslope. The coldest ground surface temperature for the same period was observed at the sideslope. The maximum difference between ground surface temperatures at the valley bottom and sideslope, and toeslope (#16a) and sideslope was observed in December: 5.8°C and 3.8°C respectively. Ground surface temperatures at the summit, shoulder and toeslope differed insignificantly for the period from October till February.

During the period from March till August ground surface temperatures at the valley bottom and toeslope (both #16 and #16a) were the coldest, but were the warmest at the sideslope.
Figure 4.8  Mean monthly ground surface temperature along Washington Creek slope, 2001-2002

Figure 4.9  Mean monthly ground surface temperature along Babe Creek slope, 2001-2002
Of interest is the considerable difference in ground surface temperatures at sites 16 and 16a, that are approximately 3 m apart at the toeslope and do not differ in vegetation or moisture content. The maximum value of mean annual ground surface temperature (MAGST) along the slope was observed at site #16a (+0.03°C) and the minimum at site #16 (-1.4°C) for the year of 2001-2002. Difference in MAGST at these sites was formed during the cold period; from April through August mean monthly ground surface temperatures at these sites practically coincided. This gives us the reason to believe that these differences in ground surface temperature were formed due to differences in warming effect of snow at these sites. Probably due to some microrelief features, the snow depth at #16a site was significantly higher than that at #16 site.

The fact that the ground surface temperature in the lower part of the slope (#15, #16a) was the warmest during the period from October through February can be explained by considerably higher site and near-surface soil water content and higher snow depths there compared to the other sites, which resulted in the highest warming effect of snow. During the period from March through August, the ground surface temperature in the lower part of the slope turned out to be the coldest one. This must be due to the lag in snow melt that occurs in the lower part of the slope in spring and causes cooling effect of the snow on the ground surface. In summer, relatively high water content on the surface and in the near-surface soils exerts a cooling effect on the ground due to high rates of evaporation from it.

Of interest is the fact that despite the coldest air temperature in the valley bottom during the cold period, the ground surface temperature was one of the warmest, hence air
temperature is not the leading factor in the formation of the local thermal regime of the ground.

The fact that the ground surface temperature was the lowest during the cold period at the sideslope and then becomes the highest during the warm period can be explained by low moisture content on the surface and in the near-surface soils and probably by the lower snow cover. We have some data to indicate that in the beginning of the cold period of 2001-2002 snow melted partially or completely from the surface at this site and, hence, a thinner snow pack compared to that of the other sites during the whole winter. This resulted in significant cooling of the ground in the beginning of the cold period and reduced warming effect of snow due the combination of dry surface conditions and low snow depths.

At the Babe Creek slope ground surface temperatures were measured at four sites: summit (#22), shoulder (#23), sideslope (#24) and toeslope (#25). Site moisture content is the highest at the summit.

MAGST is positive at all the sites and vary from +0.08°C to 1°C at sideslope and shoulder respectively (Table 4.4). Vegetation changes from partially open moss dominated woodland at the summit through fairly open shrub and moss dominated woodland to very open moss and shrub dominated bog at the valley bottom.

Variability of MAGST along the Babe Creek slope is shown in the Figure 4.9. In October, 2001 the ground surface temperatures along the slope were equal. From November through April the warmest temperatures were observed at the summit and at the rest of the sites down the slope, ground surface temperatures were very close.
Table 4.2  Mean annual and mean winter air temperature along Washington Creek and Nome Creek slopes, 2001-2002

<table>
<thead>
<tr>
<th>Slope</th>
<th># site</th>
<th>Part of the slope</th>
<th>MAAT, °C</th>
<th>MWAT, °C</th>
</tr>
</thead>
<tbody>
<tr>
<td>Washington Creek</td>
<td>12</td>
<td>Summit</td>
<td>-1.3</td>
<td>-13.8</td>
</tr>
<tr>
<td>Washington Creek</td>
<td>15</td>
<td>Valley bottom</td>
<td>-4.5</td>
<td>-20.3</td>
</tr>
<tr>
<td>Nome Creek</td>
<td>104p</td>
<td>Summit</td>
<td>-2.3</td>
<td>-12.1</td>
</tr>
<tr>
<td>Nome Creek</td>
<td>104c</td>
<td>Valley bottom</td>
<td>-4.1</td>
<td>-16.3</td>
</tr>
</tbody>
</table>

Table 4.3  Mean annual, mean summer and mean winter ground surface temperatures along Washington Creek slope, 2001-2002

<table>
<thead>
<tr>
<th></th>
<th>Summit</th>
<th>Shoulder</th>
<th>Sideslope</th>
<th>Toeslope</th>
<th>Toeslope</th>
<th>Valley bottom</th>
</tr>
</thead>
<tbody>
<tr>
<td>12</td>
<td>-0.16</td>
<td>-0.06</td>
<td>-0.94</td>
<td>-1.4</td>
<td>0.03</td>
<td>-0.7</td>
</tr>
<tr>
<td>MAGST, °C</td>
<td>11.2</td>
<td>10.8</td>
<td>11.7</td>
<td>9.4</td>
<td>10.2</td>
<td>10.3</td>
</tr>
<tr>
<td>MSGST, °C</td>
<td>-8.9</td>
<td>-8.6</td>
<td>-10.9</td>
<td>-9.2</td>
<td>-7.2</td>
<td>-9</td>
</tr>
<tr>
<td>MWGST, °C</td>
<td>0.03</td>
<td>15</td>
<td>15</td>
<td>15</td>
<td>15</td>
<td>15</td>
</tr>
</tbody>
</table>

Table 4.4  Mean annual, mean summer and mean winter ground surface temperatures along Babe Creek slope, 2001-2002

<table>
<thead>
<tr>
<th></th>
<th>Summit</th>
<th>Shoulder</th>
<th>Sideslope</th>
<th>Toeslope</th>
</tr>
</thead>
<tbody>
<tr>
<td># site</td>
<td>22</td>
<td>23</td>
<td>24</td>
<td>25</td>
</tr>
<tr>
<td>MAGST, °C</td>
<td>1</td>
<td>0.34</td>
<td>0.08</td>
<td>0.22</td>
</tr>
<tr>
<td>MSGST, °C</td>
<td>9</td>
<td>10.14</td>
<td>10.6</td>
<td>11.1</td>
</tr>
<tr>
<td>MWGST, °C</td>
<td>-3.8</td>
<td>-6.7</td>
<td>-7.1</td>
<td>-7.8</td>
</tr>
</tbody>
</table>

Table 4.5  Mean annual, mean summer and mean winter ground surface temperatures along Nome Creek slope, 2001-2002

<table>
<thead>
<tr>
<th># site</th>
<th>Summit</th>
<th>Shoulder</th>
<th>Sideslope</th>
<th>Valley bottom</th>
</tr>
</thead>
<tbody>
<tr>
<td>104 p</td>
<td>104</td>
<td>105</td>
<td>104 c</td>
<td></td>
</tr>
<tr>
<td>MAGST, °C</td>
<td>0.3</td>
<td>0.04</td>
<td>0.54</td>
<td>-0.36</td>
</tr>
<tr>
<td>MSGST, °C</td>
<td>8.1</td>
<td>8.54</td>
<td>9.14</td>
<td>8.74</td>
</tr>
<tr>
<td>MWGST, °C</td>
<td>-4.3</td>
<td>-4.76</td>
<td>-4.36</td>
<td>-5.46</td>
</tr>
</tbody>
</table>
In general, during the cold period (November-April) the ground surface temperature decreased from summit down the slope. MWGST varied from -3.8°C down to -7.8°C at the toeslope (Table 4.4). Decrease in the ground surface temperatures along the slope during the cold period can be explained by decrease in the site moisture from the summit to the toeslope, which resulted in decrease in the warming effect of snow in this direction.

In May, ground surface temperatures at all sites coincided. Then, during summer, the ground surface temperature at the summit became the coldest one: MSGST= 9°C and it increased down the slope up to 11.1°C at the toeslope. Increase of the ground surface temperatures from the summit towards the toeslope can be explained by decrease in the site moisture in the same direction and, hence, the decrease in the cooling effect of the surface due to evaporation.

Ground surface temperature measurements at the Nome Creek slope that were held at all sites along the slope, are represented in the Figure 4.10. The site moisture is slightly lower at the valley bottom (mesic), compared to the other sites (mesic to subhygric). Slight decrease in snow depths was observed in the direction from summit (0.7 m on average) to the valley bottom (0.6 m on average). MAGST vary between -0.36°C at the valley bottom to 0.54 at the sideslope (Table 4.5).

During the period from October through January, the ground surface temperature was the coldest in the valley bottom and was very similar at the rest of the sites. MWGST at the valley bottom is approximately 1°C colder than at the other sites. The combined
Figure 4.10  Mean monthly ground surface temperature along Nome Creek slope, 2001-2002

Figure 4.11  MAAT, MAGST and MAGT along Washington Creek slope, 2001-2002
effect of lower snow depth and lower site moisture gave slightly colder ground surface temperatures in the valley bottom.

During the period from May through August, the ground surface temperatures practically coincided, except for July, when at the summit it was slightly colder.

As the differences in ground surface temperatures were observed only during the cold period, it is obvious that the main factor that affects the thermal regime of the ground surface at these three slopes is the snow cover and water content on the surface and in the near-surface soils that magnifies the warming effect of snow. The fact that in case of Washington Creek slope, the strongest variability in MAGST was observed between sites #16 and #16a (that are both at the toeslope only 3 meters apart and have the same vegetation and surface conditions) proves it one more time.

Ground surface temperature is the parameter that is extremely sensitive to the surface and climatic conditions, and in each particular case its distribution will vary according to the microclimatic and local surface conditions. Hence, topographic gradient is an indirect factor in formation of the ground surface thermal regime. Our measurements show that variability within each site can be higher than along the entire profile.

**4.5 Effect of the topographic gradient on ground temperatures**

Ground temperatures at the depths of 0.35-0.5 m were measured at each site along all three slopes. Mean annual ground temperatures (MAGT) and active layer (AL) depths at each site are shown in the Table 4.6.
Table 4.6  Mean annual ground temperatures and active layer depths along the Washington Creek, Babe Creek and Nome Creek slopes, 2001-2002

<table>
<thead>
<tr>
<th></th>
<th>Summit</th>
<th>Shoulder</th>
<th>Sideslope</th>
<th>Toeslope</th>
<th>Toeslope</th>
<th>Valley bottom</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Washington Creek</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Depth of MAGT</td>
<td>0.45 m</td>
<td>0.4 m</td>
<td>0.4 m</td>
<td>0.35 m</td>
<td>0.35 m</td>
<td>0.35 m</td>
</tr>
<tr>
<td>MAGT, °C</td>
<td>0.43</td>
<td>0.27</td>
<td>0.78</td>
<td>-1.78</td>
<td>-2.1</td>
<td>-2.65</td>
</tr>
<tr>
<td>AL, 2002</td>
<td>0.95 m</td>
<td>-</td>
<td>0.92 m</td>
<td>0.37 m</td>
<td>0.37 m</td>
<td>0.39 m</td>
</tr>
<tr>
<td><strong>Babe Creek</strong></td>
<td>22</td>
<td>23</td>
<td>24</td>
<td>25</td>
<td>26</td>
<td></td>
</tr>
<tr>
<td>Depth of MAGT</td>
<td>0.51 m</td>
<td>0.51 m</td>
<td>0.44 m</td>
<td>0.41 m</td>
<td></td>
<td>0.29 m</td>
</tr>
<tr>
<td>MAGT, °C</td>
<td>-0.06</td>
<td>-0.04</td>
<td>-1.4</td>
<td>-1.2</td>
<td></td>
<td>-2.14</td>
</tr>
<tr>
<td>AL, 2002</td>
<td>0.7 m</td>
<td>0.71 m</td>
<td>0.48 m</td>
<td>0.39 m</td>
<td></td>
<td>0.42 m</td>
</tr>
<tr>
<td><strong>Nome Creek</strong></td>
<td>104p</td>
<td>104</td>
<td>105</td>
<td></td>
<td>104c</td>
<td></td>
</tr>
<tr>
<td>Depth of MAGT</td>
<td>0.52 m</td>
<td>0.48 m</td>
<td>0.46 m</td>
<td></td>
<td></td>
<td>0.53 m</td>
</tr>
<tr>
<td>MAGT, °C</td>
<td>0.51</td>
<td>0.1</td>
<td>0</td>
<td></td>
<td></td>
<td>-0.5</td>
</tr>
</tbody>
</table>
At Washington Creek slope the MAGT in the upper part of the slope (summit-sideslope) was positive in 2001-2002 (Figure 4.11). It varied from 0.27°C at the shoulder up to 0.78°C at the sideslope. In the lower part of the slope the MAGT decreased from the toeslope to the valley bottom reaching the value of -2.65°C.

Considerable decrease in active layer depths in 2002 was observed in the direction from the summit (0.95 m) to the toeslope (0.37 m) and valley bottom (0.39 m).

According to the Figure 4.12, the complete freeze-up of the active layer took place only in the lower part of the slope: at two sites at the toeslope and valley bottom. Ground temperatures at the summit, shoulder and sideslope of Washington Creek slope stayed very close to zero throughout the entire cold period; mean monthly ground temperatures did not drop lower than -0.5°C.

At the Babe Creek slope the MAGT at the summit and shoulder of the slope was very close to zero; MAGT decreased from the sideslope (-1.4°C) to the valley bottom (-2.14°C) with temperatures at the sideslope and toeslope being practically equal (Table 4.6).

Decrease in active layer depths was observed from the summit (0.7 m) to the valley bottom (0.42 m) with minimum values at the toeslope (same as at Washington Creek).

According to the Figure 4.13, complete freeze-up of the active layer definitely took place at the valley bottom, toeslope and sideslope. It is hard to say whether the complete freeze-up happened in the shoulder of the slope due to warm ground
Figure 4.12 Mean monthly ground temperature along Washington Creek slope, 2001-2002

Figure 4.13 Mean monthly ground temperature along Babe Creek slope, 2001-2002
temperatures during 2001-2002: as the MAGT was very close to zero (-0.04°C) and mean monthly ground temperatures at this site did not drop lower than -1°C.

At the Nome Creek slope MAGT changed from positive values (0.51°C) at the summit through values very close to zero at the shoulder and sideslope to -0.5°C at the valley bottom (Table 4.6).

According to the Figure 4.14, the complete freeze-up of the active layer took place only in the valley bottom. At summit, sideslope and shoulder ground temperatures were very warm; mean monthly ground temperature there did not drop lower than -0.7°C during the entire cold period.

The complete freeze-up took place in the lower part of the slopes. Decrease in the active layer depths was observed in the direction from the summits down the slopes. In the upper part of the slopes very warm ground temperatures were observed.

During the period from October till February at Washington Creek slope, the coldest ground temperatures (at the depth of approximately 0.5 m) were observed at the valley bottom and toeslope (#16a), while the ground surface temperatures were the warmest ones at these sites.

At the Babe Creek and Nome Creek slopes, maximums and minimums of the ground surface and ground temperatures coincided, being the warmest at the summits and the coldest at the valley bottoms of these slopes.

4.6 Influence of the slope aspect on ground surface temperatures

In order to evaluate the influence of slope aspect on ground surface and ground temperatures, a comparison of the ground surface and ground thermal regime between
Figure 4.14  Mean monthly ground temperature along Nome Creek slope, 2001-2002
Washington Creek and Babe Creek slope has been made. These slopes occupy approximately the same position in elevation but represent opposite aspects - S-SE (Washington Creek) and N-NW (Babe Creek).

Mean annual values of the ground surface temperature were colder along all the slope at Washington Creek compared to the Babe Creek with maximum difference at site #16 at the toeslope - 1.6°C.

Comparison between ground surface temperatures at summit, shoulder, sideslope and toeslope at Washington Creek and Babe Creek respectively are shown in the Figures 4.15 A-D.

According to Figures 4.15 A-D, during the cold period (October-March) ground surface temperatures were considerably colder at Washington Creek (S-SE) compared to Babe Creek (N-NW) at summit, shoulder and sideslope. At Washington Creek toeslope, the ground surface temperature was measured at two sites: #16 and #16a. During the cold period, ground surface temperature at site #16 was colder and at site #16a – warmer than the ground surface temperature at the toeslope of the Babe Creek.

Mean winter ground surface temperatures (MWGST) along the Washington Creek slope were considerably colder than at Babe Creek slope, except for site #16a, where the MWGST was slightly warmer. The maximum difference in MWGST was observed at the summit – 5.1°C.

The distribution of ground surface temperatures during summer was more complicated. Mean summer ground surface temperatures (MSGST) at the summit, shoulder and sideslope at Washington Creek slope were considerably warmer than at the
Figure 4.15 A

Figure 4.15 B
Figure 4.15 C

Figure 4.15 D

Figure 4.15  Effect of the slope aspect on the ground surface temperatures (2001-2002) along the slope: A-at the summit, B-at the shoulder, C-at the sideslope and D-at the toeslope
Babe Creek slope. At the lower part of the slope MSGST were considerably colder than at Babe Creek.

In general the distribution of the ground surface temperatures along these two slopes with various aspects differ significantly. The distributions of MAGST, MWGST and MSGST along the Babe Creek slope (Figures 4.16-4.18) are represented by smooth curves, that gently decrease in terms of mean annual and mean winter values and increase in terms summer means from the summit to the valley bottom.

The distribution of the mean annual, mean winter, and mean summer ground surface temperatures along the Washington Creek slope have a form of broken line with drastic changes in the ground surface temperature at the sideslope or toeslope parts (Figures 4.16-4.18).

During the year 2001-2002, some dependence of ground surface temperatures on slope aspect is observed at the upper part of the slopes (summit, shoulder and sideslope), when during the cold period ground surface is warmer at north-facing slope and during the warm period it is warmer at south-facing slope. This can be explained by partial or complete melting of the first snow from the surface on south-facing slopes that firstly gives significant cooling of the ground in the beginning of the cold period and secondly results in lower snow depths and hence lower warming effect of the snow on the south-facing slopes compared to the north-facing ones. So, slope aspect is an indirect factor that influences the thermal regime of the ground surface through the snow cover distribution and characteristics. During the warm period, snow is melting later on the north-facing
Figure 4.16 Distribution of the mean annual ground surface temperature along the slopes at Washington Creek and Babe Creek, 2001-2002

Figure 4.17 Distribution of the mean winter (2001-2002) ground surface temperatures along the slopes at Washington Creek and Babe Creek
Figure 4.18 Distribution of the mean summer (2002) ground surface temperatures along the slopes at Washington Creek and Babe Creek
slopes and the amount of incoming solar radiation is higher on the south-facing slopes, which results in higher warming of the ground surface compared to north-facing slopes.

At the toeslope and probably valley bottom, the distribution of the ground surface temperatures is much more complicated and probably is dependent on a combination of some local factors. More detailed and longer observations are needed for an explanation of this phenomenon.

4.7 Influence of the slope aspect on ground temperatures

At summit, shoulder and sideslope the active layer is deeper at Washington Creek (south-facing slope) and at toeslope and valley bottom, the active layer depth is equal at both slopes.

MAGT are higher at Washington Creek (south-facing slope) at summit, shoulder and sideslope. At the toeslope and valley bottom, MAGT is lower at Washington Creek (S-F slope) (Figures 4.19 A-E).

Complete freeze-up of the active layer took place at Washington Creek from valley bottom to toeslope and at Babe Creek from valley bottom to sideslope and even probably at the shoulder (Figures 4.19 A-E).

In the upper part of the slope, ground temperatures are warmer on the south-facing slope; complete freeze-up of the active layer is observed on the north-facing slope. In the lower part of the slopes, the complete freeze-up of the active layer happens at both slopes but the ground temperatures are colder at south-facing slope.
Figure 4.19 A

Figure 4.19 B
Figure 4.19 C

Figure 4.19 D
Figure 4.19 E

Figure 4.19 Effect of the slope aspect on the ground temperatures (2001-2002) along the slope: A-at the summit, B-at the shoulder, C-at the sideslope, D-at the toeslope and E-in the valley bottom.
4.8 Conclusions

- Mean annual air temperatures in the Fairbanks region varied from -4.5 to 0.1°C in 2001-2002. Mean annual ground surface temperatures in the region differed from -3.9 to 4.3°C. Differences in mean annual air and ground surface temperatures are formed mainly during the winter period. Mean annual ground temperatures vary from -4.9 to 1.7°C. Active layer depths vary from 0.3 to more than 0.9 m within the region. The only observed pattern in the distribution of the air, ground surface and ground temperatures in the Fairbanks region is along the local topo-sequences.

- Strong air inversions were observed along the local topo-sequences during the cold period (up to 8°C difference in mean monthly air temperatures between the summit and the valley bottom with the latter being colder). During the warm period, air temperatures at the summit and at the valley bottom are equal.

- The ground surface temperature regime along the topo-sequences, during the cold period, is determined by the combined effect of the snow cover and water content on the surface and in the near-surface soils. The higher snow depths and higher water content result in the higher ground surface temperatures.

- Some local events, like partial melting of the first snow on some parts of the topo-sequences, can result in significant changes in the thermal regime of the ground surface, causing the cooling of the ground surface and maintaining the lower snow-depths and hence colder ground surface temperatures through the entire cold period.

- During the snow-free period, the ground surface thermal regime is determined by the water content at the surface and in the near-surface soils. The higher the water
content, the lower the ground surface temperatures due to the heat loss because of the evaporation from the ground surface.

- Variability of the ground surface temperatures within each site (for example sites #16 and #16a at the Washington Creek) can be stronger than along the whole topo-sequence).

- The complete freeze-up of the active layer took place on the lower part of the slopes. Decrease in the active layer depths was observed in the direction from the summit to the valley bottom. In the upper part of the slopes, very warm ground temperatures were observed compared to the lower part of the slopes.

- The influence of the slope aspect on the ground surface temperatures works through the solar radiation during warm period and through the combined effect of the solar radiation and snow cover during the cold period. During the cold period the ground surface is generally warmer (except for #16a) at the North-facing slope because of the melting of the first snow at the south-facing slope and hence cooling of the ground surface. During warm period, the ground surface is generally warmer at the South-facing slope (except for the lower part) due to the higher amount of the solar radiation incoming to the surface.

- In the upper part of the slopes, ground temperatures are warmer at the south-facing slope. The complete freeze-up of the active layer was observed only at the north-facing slope. In the lower part of the slopes the complete freeze-up of the active layer happens at both slopes, but the ground temperatures are colder at the south-facing slope.
CHAPTER 5

Changes in Air and Ground Temperature Regime across Various Landscapes

To reveal the effect of climatic and surface parameters, such as vegetation cover, snow cover, soil properties (primarily water content) on the ground thermal regime, the Smith Lake profile was chosen. The map of Smith Lake measurement sites is shown in the Figure 5.1. It consists of five sites that belong to different landscapes, including typical coniferous forest, tussocky forest-tundra, transition zone between the last two and typical deciduous forest. Hence, these five sites represent various microclimatic and permafrost conditions, vegetation and soil characteristics and differ slightly in absolute elevation and slope aspect.

Measurements of air, snow surface, ground surface, ground temperatures, water content at different depths, snow and active layer (AL) depths were made at these sites regularly for different periods between 1996 and 2003. Most of these sites originally were established by Prof. Chien-Lu Ping and his colleagues.

5.1 Profile description

The Smith Lake-1 site (SL-1) (Figure 5.2A) is situated in the typical coniferous forest dominated by white spruce. The shaded area of the site is approximately 50%. Distance between trees varies in between 1-4 m. The prevailing ground surface cover is the feather moss (Figure 5.2B). Moss occupies about 80% of the ground surface and the remaining 20% is covered with grass. Smith Lake-1 site represents the upper part of the gentle slope with southern aspect.
Figure 5.1 Schematic display of the Smith Lake profile
Figure 5.2 A

Figure 5.2 B

Figure 5.2  A–the Smith Lake-1 site, B–ground surface cover at the Smith Lake-1 site
Hourly measurements of air temperature have been held at SL-1 site for the period from mid October, 2001 till present using HOBO sensor that was placed inside the radiation shield to avoid heating from direct solar radiation. The snow surface temperature was measured hourly for the same period with two HOBO sensors approximately 5 m apart. Both sensors lay in the shadowed place under the black spruces at the distance approximately 30 cm to the north-east from the trunk, avoiding the direct solar radiation. During the period when the snow is present on the ground, sensors were placed right on the snow surface measuring the snow surface temperature. Sensors were covered with thin layer of white paint in order to maintain the same albedo as the snow cover. During warm period, loggers measured the ground surface temperature at the depth of approximately 0.01-0.02 m. Hourly measurements of the ground surface temperature at the depth of 0.01-0.02 m throughout the entire year were carried out with Optic Stowaway sensor starting from October, 2001 till present. Hourly measurements of ground temperatures were collected, for the period from October, 1996 till present starting from the surface (0.02 m) to the depth of 0.77 m at 10-15 cm intervals, by an MRC sensor. Water content in soils at the depths of 0.2, 0.4, 0.6 and 1m was measured hourly for the period from October, 1996 till now with VITEL sensors. MRC and Optic Stowaway loggers were placed in a small open area.

Snow depths at the SL-1 site were measured near each sensor manually after each snowfall during two winters – 2001/2002 and 2002/2003. According to measurements conducted in August, 2002, active layer (AL) depth was 0.68 m on average at the SL-1.
The Smith Lake-3 site (SL-3) (Figure 5.3A) is situated in the transition zone between the typical coniferous forest and tussocky forest-tundra. The prevailing ground surface covers are the feather moss and the sphagnum moss that cover approximately 60% of the site (Figure 5.3B). About 30% of the Smith Lake-3 site is covered with dwarf shrubs and approximately 20% is occupied with black spruce. This site represents the lower part of the gentle slope with southern aspect.

Surface water conditions can be characterized as wet, as during the snow melt a 5-7 cm layer of ice forms in the moss. In late spring and summer, standing water in surface depressions is observed.

Hourly air temperature was measured for the period from mid October, 1997 till present with L-107 sensor. Hourly snow surface temperature was measured starting from October, 2001 till present with four HOBO sensors under the black spruce, avoiding the direct solar radiation, at the distance about 30 cm to the north-north-east from the trunk and 10-15 cm from one another. During the snow-free period these HOBO sensors were used for measurements of the ground surface temperature at the depth of 0.02 m. Hourly ground surface temperatures at the depth of 0.01 and 0.07 m were measured for the same period of time with L-107 sensors. Hourly ground temperatures at several depths starting at 0.25 down to 1 m with the intervals 10-25 cm was measured from October, 1997 till present with MRC sensor. Hourly water contents in soils for the same period were measured by six VITEL sensors, at the depths of 0.35, 0.55 and 0.9 m, with two sensors at each depth. MRC and L-107 were situated in the open part of the SL-3.
Figure 5.3 A

Figure 5.3 B

Figure 5.3  A-the Smith Lake-3 site, B-ground surface cover at the Smith Lake-3 site
Measurements of snow depth at Smith Lake-3 site were made during the winters of 2001/2002 and 2002/2003 after each snowfall near each logger and all over the site. Active layer depth measured with probe in August 2002 was 1.04 m near MRC and 0.71 m near Hobo sensors.

The **Smith Lake-4** site (SL-4) (Figure 5.4A) represents a typical tussocky forest-tundra. Tussocks are about 30-50 cm in height. Approximately 30% of the ground surface is covered with feather moss, and the rest is covered with sedges (Figure 5.4B). Higher level vegetation is represented only by the sparse black spruce with undergrowth of the white spruce. SL-4 site can be considered as very wet, as the water is standing in between tussocks throughout the entire warm period and into the early winter.

Hourly measurements of air temperature are held for the period from November, 2001 till present with L-107 sensor. Hourly snow surface temperature was measured from mid-October, 2002 till present with L-107 sensor. Measurements of the ground surface temperature in between tussocks with hourly interval at the depth 0.02 m were held with Optic Stowaway sensor from October, 2001 till July, 2002 and with L-107 sensor for the period mid-October, 2002 till present. Hourly measurements of ground temperature starting from 0.02 with the interval every 10-15 cm to the depth of 1 m were taken with MRC sensor for the period from mid-January, 1999 till present.

Snow depth measurements were made during the 2002/2003 winter after each snowfall on the tussocks and in between them. According to our measurements, AL depth was approximately 0.5 m in September, 2002.
Figure 5.4 C

Figure 5.4 A-the Smith Lake-4 site, B-ground surface at the Smith Lake-4 site, C-openings through the snow between tussocks at the Smith Lake-4 site
The **Smith Lake-2** site (SL-2) (Figure 5.5A) is the transition zone between the tussocky forest-tundra area and coniferous forest. About 50% of the ground surface is occupied by feather moss and sphagnum moss and lichens and the rest is covered with sedges (Figure 5.5B). Black spruce occupies approximately 30% of the site area. This site represents the lower part of the gentle slope of northern aspect.

Hourly snow surface temperature measurements were collected for the period from October, 2001 till present with two HOBO sensors about 5 m apart. Sensors were placed under the spruce, avoiding the direct solar radiation. One of them is 10 cm from the trunk to the east, the second one – 20 cm from the trunk to the north-east. During the snow-free period, these HOBO sensors were used for measurements of the ground surface at the depth of 0.02 m. Hourly ground temperatures at the depths of 0.3, 0.52, 0.85 m and water content in soils at the depths of 0.2, 0.4, 0.65, 0.85 m were measured from mid-September, 1998 till mid-September, 2001 with L-107 and VITEL sensors respectively. MRC probe was placed in an open part of Smith Lake-2 site in 2000. According to our measurements, AL in September, 2002 reached 0.6 m on average.

The **Potato Field** site (PF) (Figure 5.6) differs significantly from the other sites of the profile. It is situated in the typical deciduous forest. Among the ground surface cover, only leaf litter is present on the ground, there is no moss or grass cover present. The prevailing tree types are birch, aspen and shrubs; spruce trees are practically absent. Trees and shrubs cover approximately 70% of the ground surface at Potato Field site.

Hourly air temperature measurements were held for the period from mid-October, 2001 till present with one HOBO sensor that was placed inside the radiation shield. Snow
Figure 5.5 A

Figure 5.5 B

Figure 5.5  A-the Smith Lake-2 site, B-ground surface cover at the Smith Lake-2 site
Figure 5.6  Ground surface cover at the Potato Field site
surface temperature was measured on an hourly interval starting from mid-October, 2001 till present with three HOBO sensors that were placed under a birch-tree 1 m to the north from the trunk approximately 20 cm apart. During the snow-free period, these HOBO sensors measured the ground surface temperature at the depth of 0.01-0.02 m. Hourly measurements of the ground surface temperature at the depth of 0.01-0.02 m were held with Optic Stowaway sensor from mid-October, 2001 till present.

Snow depths were measured during the winters of 2001/2002 and 2002/2003 after each snowfall near each sensor. The permafrost surface was not reached with a probe 1.3 m long. It appears that permafrost is absent at this site, at least in the upper few meters.

### 5.2 Air temperature

Variations in mean monthly air temperature along the profile are shown in the Figure 5.7. Mean monthly air temperatures along the Smith Lake profile give good correlation with that from the Fairbanks International Airport weather station (Figure 5.7).

The difference in air temperatures between sites is mostly pronounced during the cold period. The warmest air temperatures during both winters were observed at the PF and the coldest at SL-4 and SL-2. During the winter of 2001-2002, the mean winter air temperature (MWAT) at PF was 1.5°C warmer than at the SL-1 site and 2.3°C warmer than at the SL-3 and SL-4 (Table 5.1). Here and after, the mean winter temperatures are calculated as average over December through February. During the winter of 2002-2003, the air temperatures in general were warmer by approximately 3°C at all sites, compared
Figure 5.7  Mean monthly air temperatures along the profile and at the Fairbanks International Airport weather station (FAI) (data on FAI adopted from NWS)
Table 5.1 Comparison between mean winter and mean summer air temperatures along the profile

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to the winter of 2001-2002. The differences in MWAT for the winter of 2002-2003 were 1°C, 0.8°C, 1.5°C and 1°C between PF and SL-1, SL-3, SL-4 and SL-2 sites respectively.

The difference in mean summer air temperatures (MSAT) did not exceed 0.6°C, so may be considered negligible.

The minimum observed daily air temperature for the winter of 2001-2002 was between -38.4°C (at PF) and -40°C (at SL-1); and between -36.2 (PF) and -38.3°C (at SL-1) for the winter of 2002-2003. The maximum for the summer of 2002 was +20-21°C for all sites. The range of seasonal variations in air temperature is practically the same among sites and is about 30°C (varying from 29.6 to 30.03°C).

5.3 Snow surface temperature

The snow surface temperature for the last two winters and the ground surface temperature for the summer are shown in the Figure 5.8. In general, the snow surface temperature is very much alike for all five sites, except for slightly lower values at SL-2 (up to 3.5°C colder than at the other sites) in March, 2002. The difference in mean monthly snow surface temperatures does not exceed 1.5°C.

Very slight variability in snow surface temperatures was observed within each site (between the sensors), so it can be considered negligible. Results on the snow surface and air temperature measurements made by three and one HOBO sensors respectively at the PF site are represented in the Figure 5.9. Snow surface temperatures measured by three sensors gave a good correlation. During the period October-December, the air temperature coincides with three snow surface temperatures, then from December till March some colder air temperature peaks are observed and there is a short period in
Figure 5.8 Mean monthly snow or ground (during the snow-free period) surface temperatures along the profile

Figure 5.9 Daily snow surface (3 sensors) and air temperature at the Potato Field site
March when the air temperature is slightly warmer than the snow surface temperatures, but still below 0°C. In general, the snow surface temperature is either equal or warmer than the air temperature at the Potato Field site.

5.4 Snow surface temperature vs. air temperature

According to our measurements, mean winter snow surface temperature (MWSST) was higher than the mean winter air temperature (MWAT) at all sites during the last two winters, except for the SL-2 site during the winter of 2002/2003, when MWAT and MWSST were practically equal. This difference varied from 1.5°C for PF during the winter of 2002/2003 and up to 5.4°C at SL-3 site during the winter of 2001/2002 (Table 5.2).

Monthly relations between snow surface and air temperatures are shown in the Figures 5.11 through 5.15. As we can see from the graph for both years, the snow surface temperature at the SL-1 site (Figure 5.11) is equal to the air temperature in November. During the winter months, it is considerably warmer than the air temperature and in March and April it equals the air temperature again.

At the SL-3 site (Figure 5.12), mean monthly snow surface temperatures are warmer for the entire cold period: November-February. In March and April the snow surface and air temperatures practically coincide.

At the SL-4 site (Figure 5.13), the relationship between the air and the snow surface temperatures is more complicated: in November and January snow surface temperature equals the air temperature, in December and February it is warmer than the
Figure 5.10  Warming effect of snow along the profile

Figure 5.11  Mean monthly air, snow/ground (during the snow-free period) surface, ground surface and ground temperatures at the Smith Lake-1 site
Table 5.2  Comparison of mean winter snow surface and mean winter air temperatures along the profile

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Figure 5.12  Mean monthly air, snow/ground (during the snow-free period) surface, ground surface and ground temperatures at the Smith Lake-3 site

Figure 5.13  Mean monthly air, snow/ground (during the snow-free period) surface, ground surface and ground temperatures at the Smith Lake-4 site
Figure 5.14  Mean monthly air, snow/ground (during the snow-free period) surface, ground surface and ground temperatures at the Smith Lake-2 site.

Figure 5.15  Mean monthly air, snow/ground (during the snow-free period) surface and ground surface temperatures at the Potato Field site.
air temperature and then in March and April, it is slightly colder. In mean winter values, the snow surface temperature was 1.8°C warmer than the air temperature in 2002/2003.

At the SL-2 site (Figure 5.14), the relationship between air and snow surface temperatures changes during the cold period. In mean winter values for 2002/2003, the air and snow surface temperatures are practically equal.

At the PF site (Figure 5.15), the snow surface temperature is warmer than the air temperature during both winters and slightly colder in March and April of both years.

So, in general, the snow surface temperature during the cold months is either warmer or equal to that of the air, except for the spring period (March and April) when it is slightly colder than the air at all sites during both years. Slightly colder snow surface temperature during the spring months can be explained by the fact, that when the air temperature raises above zero, the snow starts to melt from the surface and in this way keeps the snow surface temperature close to zero due to phase transition. The sublimation processes on the snow surface can also be the reason for the slightly colder snow surface temperatures in spring. With the sufficient amount of the solar radiation incoming to the ground surface during the day, (even with the air temperatures below zero) this process can take place, decreasing drastically both the snow depth and the snow surface temperature (due to the consumption of energy on phase transition).

5.5 Warming effect of snow

In the Figure 5.10, the differences between snow surface and ground surface temperatures for each site are shown. The difference is interpreted as the warming effect
of snow. The warming effect of snow during the winter of 2001/2002 was the most pronounced at the SL-3 and in spring of 2002 at the SL-2. During the winter of 2002/2003, the strongest warming effect of snow was once again at the SL-3 site and the second strongest was at the SL-2 site. The warming effect of snow at the SL-3 site reached as high as 16°C in December, 2001 and even 19.3°C in January, 2003. At the SL-2 site, the warming effect of snow reached 17.1°C in January, 2003. Such strong warming effect of snow at the SL-3 and SL-2 sites can be explained by the thickest snow cover among other sites of the profile practically through the entire winter of 2001-2002 at both sites and for the SL-2 site during the winter of 2002/2003. Slightly higher warming effect of snow at the SL-3 site compared to that at the SL-2 is probably due to the higher water content on the surface and in near-surface soils, as thermal properties and AL water content can have a significant influence on the warming effect of the snow cover (Lachenbruch, 1959; Romanovsky, 1987). The minimum warming effect of snow was observed at the PF site during both years. It did not exceed 8°C, which was probably due to the lowest snow depths during both winters and dry conditions at the ground surface and in the near-surface soils.

Despite the highest snow depths and the highest water content at the surface and in the near-surface soils at the SL-4, compared to the other sites of the profile, the warming effect of snow was not the highest one. This can be explained by significant cooling of the ground through the openings at the sides of tussocks and through the ice layers between the tussocks.
The warming effect of snow differed significantly between springs of 2002 and 2003. For April, 2002 the warming effect of snow differs between 1 and 3°C for the SL-3 and PF sites respectively, but in April, 2003 it is about 1.5°C for the SL-3 site, 1.4°C for the SL-2 site. It is practically absent (0.07°C) at the SL-1. At the SL-4 and PF sites, snow cover already exerts the cooling effect on the ground surface. This is simply due to the fact that the air temperatures in April for the SL-4 and PF site were already above 0°C in April.

In general, the warming effect of snow during the winter of 2002/2003 was considerably higher, compared to that of 2001/2002 due to the fact that snow depths were approximately twice as high during the winter of 2002/2003.

5.6 Ground surface temperature

The ground surface temperature was measured at the depth of 0.02 m from the moss cover surface. Variability of ground surface temperatures along the profile is shown in the Figure 5.16. The mean winter ground surface temperature (MWGST) for 2001-2002 (Table 5.3) was the coldest at the SL-4 site -13.3°C. Extremely warm ground surface temperatures were observed at the SL-3 site: the MWGST was -4.2°C. For the year of 2001-2002, the MWGST decreased in order: SL-3 – SL-1 – PF – SL-4.

For the winter of 2002-2003, MWGST were significantly warmer at all sites. Once again, extremely warm ground surface temperatures were observed at the SL-3 site (MWGST= -2°C). The coldest ground surface temperature was at the PF (MWGST=-7.5). For this winter, the MWGST decreased in order: SL-3, SL-2, SL-4, SL-1, PF.
Figure 5.16  Mean monthly ground surface temperatures along the profile

Figure 5.17  Snow depths along the profile during the 2001/2002 winter
Table 5.3 Comparison between mean winter and mean summer ground surface temperatures along the profile

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During the summer of 2002, the ground surface temperatures were very close at SL-1, SL-2 and SL-3 sites: mean summer ground surface temperatures (MSGST) were 10.1°C, 10.3°C and 10.5°C respectively. MSGST at PF was slightly warmer: 11.5°C. Cold ground surface temperatures during this summer were observed at the SL-4 site: 5.1°C.

During the warm period (from May to September), the ground surface temperature at PF is the warmest. This might be explained by the highest air temperature at the Potato Field site, among the others, and dry surface conditions that enable the ground to warm faster due to absence of significant evaporation.

During the winter, ground surface temperature at PF was one of the coldest (after the SL-4) in 2001-2002 and the coldest one during 2002-2003, despite the fact that the air temperature at PF site was the warmest during both winters.

As we can see from Figures 5.17 and 5.18, the snow depth at PF site was the lowest among the other sites during both winters. Also of interest is the fact that in the mid-November, 2002 the snow depth at PF suddenly decreased by more than 5 cm probably due to the thaw that took place during that time and from that point stayed the lowest throughout the entire winter. This resulted in a decrease in the warming effect of snow and in the coldest temperatures of the ground surface at the PF. Another cause of the decrease in warming effect of snow is dry surface conditions at the PF, as all the rest of the sites are characterized by moderate or high water content near the surface. So, low snow depths and dry surface conditions resulted in significant cooling of the ground surface during both years of measurements at the PF.
Figure 5.18  Snow depths along the profile during the 2002/2003 winter
Extremely warm ground surface temperatures were observed at the SL-3 site during both winters, despite the fact that the air temperatures were the coldest here during both winters. Warm ground surface temperatures can be explained by the high water content at the surface and near surface soils and thick snow cover during both winters, as high water content at the surface and in the near-surface soils results in increase of the warming effect of snow.

The data on ground surface temperatures for the SL-2 site is only available for the period from March, 2002 till present. For this period, the ground surface temperature at the Smith Lake-2 site turned out to be the second warmest (after the SL-3 site). Such warm ground surface temperature is probably due to the thick snow cover that reached 0.5 m in the winter of 2002/2003 and due to moderate water content in the surface and near-surface soils.

The SL-1 site occupies intermediate position in ground surface temperatures between other sites throughout the entire year. The difference in MWGST with the SL-3 site was 5.5°C in 2001-2002 and 3.2°C in 2002-2003. The reason for that can be much lower water content on the surface and in the near-surface soils at Smith Lake-1 and, therefore, the warming effect of snow was not so highly pronounced here, compared to SL-3. Also, the snow cover was almost the same at both sites for the 2002/2003 winter and it was much lower at SL-1 compared to SL-3 for 2001/2002 winter. In summer, the ground surface temperature at SL-1 and SL-3 were practically the same.

Distribution of ground surface temperatures at the SL-4 site is much more complicated, compared to the other sites. In November, 2001 the ground surface
temperature is equal to that of SL-3 – the warmest one, then in December it drastically decreases by almost 9.5°C and stays the coldest among other sites till November, 2002, when the ground surface temperatures practically coincide at all sites. In December, 2002 and January, 2002 it stays the warmest compared to other sites. Then, once again, we observe the drastic decrease in the surface temperature and in February it becomes the coldest one and stays like that till April, 2003. Warm temperatures of the ground surface during November, 2001 were probably due to the standing water between tussocks that prevented surface from cooling, as all the heat is consumed on phase transitions. Significant cooling of the ground in the beginning of December can be explained by the fact that firstly, all water between the tussocks is already transformed into the ice, which has high thermal conductivity and secondly by the presence of openings through the snow cover at the sides of the tussocks (Figure 5.4C) that allow the convective heat flow from the cold atmosphere and snow surface. So, despite the very thick snow cover (thickest among all the sites) – up to almost 0.6 m during the year of 2002/2003 (Figure 5.18), the ground surface experiences significant cooling.

A similar occurrence happened in the winter of 2002/2003, although the complete freezing of the water between tussocks took place much later (in the end of December – beginning of January) and the considerable ground surface cooling started only in January.

During the warm period, ground surface temperature at SL-4 site stays as the coldest one, the temperature does not rise above 5.6°C. The difference in MSGST between SL-4 and other sites was 5-6°C. One should keep in mind that the ground
surface temperature at the SL-4 site was measured between tussocks (in the shaded area) and the ground surface on the top of the tussocks can be considerably warmer. Such cold ground surface temperatures may also result from extremely wet surface conditions that cause significant cooling of the surface due to high rates of evaporation.

Strong spatial variability of ground surface temperatures was observed among the sites during both years. The maximum difference in MWGST was 9.18°C between the SL-3 and SL-4 in 2001-2002 and 5.5°C between the SL-3 and PF in 2002-2003. The maximum difference in MSGST was 6.34°C between PF and SL-4 site. This considerable variability can be explained by combined effect of snow cover, surface cover characteristics and especially water content on the surface and in near-surface soils.

5.7 Ground temperatures

Variability of the ground temperatures at maximum depths of measurements (0.85-1 m) along the profile is shown in the Figure 5.19. Ground temperature from the surface to the depth of approximately 1 m was measured at the SL-1, SL-3 and SL-4 sites for the period from 1997, 1998 and 1999 respectively till present. At the SL-2 site the ground temperature was measured for the period from 1998 till 2001 at the depths of 0.3 and 0.85 m and for the period from 2001 till present from the surface to the depth of 1 m.

At the SL-1 site (Figure 5.20) MAGT (mean annual ground temperatures) at the depth of 0.5 m were negative for the entire period of observations and varied from -0.5 to -1.03°C. For the period from 1997 till 2002, active layer depths varied between 0.6 and 0.65 m with its maximum in September of 1997 and 1998 (Figure 5.21).
Figure 5.19 Comparison of the ground temperatures at maximum depth of measurements (0.7-0.95 m) along the profile

Figure 5.20 Comparison between mean annual ground temperatures at the depth of 0.5 m along the profile
Figure 5.21  Vertical temperature profiles at the Smith Lake-1 for September and March, 1997-2003
If we compare the mean monthly temperatures of the ground in March, when the coldest ground temperatures are observed (when cooling of the ground reaches the maximum depth of observations) and in September, as the month with maximum thaw depth, we can conclude that the complete freeze-up of the active layer took place every year during the period of observation, as the interannual variations in mean monthly March temperatures at the depth of 0.7 m are in the range from -2.0 (in 2003) to -3.7°C (in 1999). During March, the unfrozen water content in soils decreased from 10% at the depth of 0.4 m through 7% at 0.6 m to 3% at 1 m (Figure 5.22).

According to measurements at the SL-3, the active layer depth in September, 2002 was 1.04 m near the MRC sensors. According to Figures 5.23-5.25, the complete freeze-up of the active layer took place during the years of 1999 and 2002, and, probably, in 2001. MAGT at the depth of 0.5 m was -0.22°C in 1998-1999 and -0.41°C in 2001-2002.

MAGT at the depth of 0.5 m was +0.43°C in 1999-2000. During March of 1998, 2000 and 2003 years the "zero curtain" could still be observed at 0.8 m (1998) and 0.6 m (in 2000 and 2003) (Figure 5.24).

In 2001, using the temperature data, it is hard to say, whether the complete freeze-up did take place, as the maximum depth of the ground temperature measurements was only 0.8 m that year. The MAGT at the depth of 0.5 m was extremely close to zero (+0.06°C). According to Figure 5.25, it is more likely that the complete freeze-up of the AL happened in 2001.
Figure 5.22  Water content in soils at Smith Lake-1 and Smith Lake-3 (for Smith Lake-1 curves for 2000 and 2003 practically coincide)
Figure 5.23 Mean monthly ground temperature at the Smith Lake-3 for 1997-2003
Figure 5.24  Vertical temperature profiles at the Smith Lake-3 for September and March, 1998-2003
Figure 5.25  Temperature field dynamics within the active layer and near-surface permafrost at the Smith Lake-3, 1997-2003
During the whole period of measurements in March the maximum water content was observed in the middle of the AL (Figure 5.22).

The ground temperature at the **SL-4** site (Figure 5.26) is significantly colder than that at SL-1 and SL-3. Values of MAGT at the depth of 0.5 m are available only for 2000-2001, 2001-2002 and 2002-2003 years and are -3.04, -4.72 and -2.34°C respectively (Figure 5.20). Considerable difference in MAGT at the depth of 0.5 m is observed between sites: for example, for 2001-2002 the difference between SL-4 and SL-1 was 3.79°C and even 4.31° between SL-4 and SL-3. Considerable interannual variations that range from -12.2 (in 1999) to -6.8°C (in 2000) in mean March ground temperatures at the depth of 1 m were observed (Figure 5.26). The AL depths ranged from 0.45 m (in 1999) to 0.6 m (in 2001); the complete freeze-up of the active layer took place every year from 1999 to 2003 (Figure 5.26).

Such cold ground temperatures and early freeze-up dates (for example, in November) that are observed at the SL-4 site are not typical for the Fairbanks region. We believe that such cold ground thermal regime can allow the formation and preservation of the ice wedges at the SL-4 site.

At the **SL-2** site the ground temperature distribution with depth during the period 1998-2003 is shown in the Figure 5.27. According to this graph, AL reached approximately 0.5 m in 1999, then it was about 0.6-0.7 m during 2000 and 2001 (Figure 5.27) and in 2002 AL was 0.65 m (Figure 5.28).

According to Figure 5.27, the freeze-up happened every year during 1999-2001. During this period, the coldest mean monthly ground temperature -7.05°C at the depth of
Figure 5.26  Vertical temperature profiles at the Smith Lake-4 for September and March, 1999-2003
Figure 5.27  Mean monthly ground temperature at the Smith Lake-2 for 1998-2003
Figure 5.28  Vertical temperature profiles at the Smith Lake-2 for September and March, 2002-2003
0.85 m was observed in March 1999. Then, in March 2000 it was -3.98°C and in 2001 - 4.65°C. The distribution of water content in soils with depth in March is shown in Figure 5.29. The behavior of the water content curves for the years 1999-2001 for March are similar to that at the SL-3, as during all three years the maximum water content in March was observed in the middle of the AL.

In the Table 5.4, comparisons between mean annual air temperature, ground surface cover (the snow surface when the snow is present on the ground and the ground surface (at the depth of 0.02 m below the moss surface) during the snow-free period) and ground (0.5 m) temperatures along the profile are shown. Mean annual temperatures are calculated for two years with the same summer period: the first year - November, 2001 through October, 2002 and the second - May, 2002 through April, 2003.

Data on MAAT and MAS/GST during the year starting from November, 2001 till October, 2002 are only available for the SL-1 and PF. For the SL-1 site the difference between MAAT and MAS/GST 0.48°C with colder air temperature and for the PF it was 0.39°C with the colder temperature of the ground surface covers.

For the second year starting from May, 2002 till April, 2003, the difference between MAAT and MAS/GST was practically absent at the SL-1 site, it was 0.65°C, 1.28°C and 1.58°C at the PF, SL-2 and SL-4 sites respectively with the colder temperatures of the ground surface covers.

So during both years difference between MAAT and MAS/GST was practically negligible at the PF and at the SL-1 site and it was slightly larger at the SL-2 and SL-4 sites, but in terms of mean annual values, it may be considered small. One should keep in
Table 5.4 Comparison between mean annual air, ground covers, ground surface (0.02 m) and ground (0.5 m) temperatures along the profile

<table>
<thead>
<tr>
<th>Nov,01-Oct,02</th>
<th>Potato Field</th>
<th>Smith Lake-2</th>
<th>Smith Lake-4</th>
<th>Smith Lake-3</th>
<th>Smith Lake-1</th>
</tr>
</thead>
<tbody>
<tr>
<td>MAAT</td>
<td>-2.51</td>
<td>-</td>
<td>-3.9</td>
<td>-4.13</td>
<td>-3.68</td>
</tr>
<tr>
<td>MAS/GST</td>
<td>-2.9</td>
<td>-4.03</td>
<td>-</td>
<td>-</td>
<td>-3.2</td>
</tr>
<tr>
<td>MAGST</td>
<td>-0.89</td>
<td>-</td>
<td>-3.91</td>
<td>1.56</td>
<td>-0.39</td>
</tr>
<tr>
<td>MAGT</td>
<td>-</td>
<td>-</td>
<td>-4.89</td>
<td>-0.43</td>
<td>-0.94</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>May,02-Apr,03</th>
<th>Potato Field</th>
<th>Smith Lake-2</th>
<th>Smith Lake-4</th>
<th>Smith Lake-3</th>
<th>Smith Lake-1</th>
</tr>
</thead>
<tbody>
<tr>
<td>MAAT</td>
<td>-0.65</td>
<td>-1.84</td>
<td>-1.94</td>
<td>-2.21</td>
<td>-1.92</td>
</tr>
<tr>
<td>MAS/GST</td>
<td>-1.3</td>
<td>-3.12</td>
<td>-3.52</td>
<td>-</td>
<td>-1.94</td>
</tr>
<tr>
<td>MAGST</td>
<td>1.05</td>
<td>1.83</td>
<td>-0.71</td>
<td>2.66</td>
<td>0.9</td>
</tr>
<tr>
<td>MAGT</td>
<td>-</td>
<td>-0.76</td>
<td>-2.34</td>
<td>0.08</td>
<td>-0.49</td>
</tr>
</tbody>
</table>
mind that this difference between MAAT and MAS/GST was formed mainly due to the colder ground surface temperatures in summer, compared to that of the air, because they were measured at the depth of 0.02 m below the moss surface.

Thermal offset depends on soil thermal properties and is usually the greatest within a peat layer (Romanovsky, 1989; Romanovsky & Osterkamp, 1995). For homogeneous semi-infinite soil system, Romanovsky and Osterkamp (1995) have shown that thermal offset is proportional to thawing index of the ground surface temperature but inversely proportional to the ratio of the thawed to frozen conductivity of soils. Figure 5.30 shows the comparison of vertical temperature profiles using mean annual values for the period from May, 2002 till April, 2003.

For calculations of the thermal offset temperatures measured at the depth of approximately 0.02 m in the moss cover were taken as the ground surface temperature; the mean annual temperature at the bottom of the active layer was calculated using the data from measurements. The depth of the active layer was measured by hand in September, 2002.

The thermal offset varied by 1.5-2 times among the sites. It was -1.4, -1.6, -2.6 and -2.8°C at SL-1, SL-4, SL-2 and SL-3 respectively. According to Osterkamp and Romanovsky (1999), these numbers are rather typical for Interior Alaska. Such high values of the thermal offset at the SL-2 and SL-3 sites can be explained by high heat turnovers due to high water content in the near-surface soils (especially at the SL-3 site) that significantly increase the difference between thermal conductivity of soils in thawed
Figure 5.29  Water content in soils at the Smith Lake-2 in March, 1999-2003
Figure 5.30  Vertical temperature profiles (mean annual values) at Smith Lake-1, Smith Lake-2, Smith Lake-3 and Smith Lake-4
and frozen state and by the presence of the peat layer on the ground surface, in which the thermal offset is the largest.

Mean annual ground temperature decreased with depth in the AL at all sites, except for the SL-1 site, where starting from the depth of approximately 0.2 m MAGT stayed constant (-0.5°C) at least till the depth of 0.6 m.
5.8 Conclusions

- Small spatial variability of air temperature (mainly during the cold period) was observed along the profile. In terms of mean winter values, the maximum observed difference between the sites was 2.3°C and 1.5°C for the winters of 2001/2002 and 2002/2003 respectively. The difference in mean summer air temperatures did not exceed 0.6°C, and so may be considered negligible. Measured data on air temperature along the Smith Lake profile correlate well with that at the Fairbanks International Airport weather station.

- Very little difference was observed in snow surface temperatures between the sites during both years. The difference in mean monthly snow surface temperatures between sites did not exceed 1.5°C.

- The difference between the mean annual air and snow surface/ground surface (during the snow-free period) temperatures with the last one being colder, was formed due to the colder ground surface temperatures in summer.

- Snow surface temperatures during the cold months were either warmer (in terms of mean winter values up to 5.4°C for SL-3 for the winter of 2001/2002) or equal to that of the air along the profile during both years.

- In spring, snow exerts a cooling effect on the ground surface when the air temperature rises above 0°C and the snow cover prevents the ground surface from warming.

- The main factors that determine the thermal regime of the ground surface during the cold period at the SL-1, SL-2 and SL-3 sites are the snow cover (that exerts the warming effect) and the water content on the surface and in the near-surface soils...
(that amplifies this effect). At the PF site the main factor, which influences the ground thermal regime is the absence of moss cover on the ground (due to the presence of leaf litter on the ground that prevents the growing of the moss cover). The warming effect of snow turned out to be the highest at the SL-3 site, (site with the high snow depths and high water content in the near-surface soils) and the minimum – at the PF site (site with the lowest snow depths and dry surface and near-surface soil conditions). The SL-4 site turned out to be an exception from this pattern, as in spite of the highest (among other sites of the profile) snow depths and water content at the surface and in the near-surface soils, the ground surface temperature was the coldest due to the existence of openings at the sides of tussocks and, hence, convective heat transfer through these openings during the first part of winter and conductive heat transfer through the ice lenses that form due to the surface water freezing, that resulted in the fast and strong cooling of the ground. Hence the vegetation cover in form of tussocks plays the leading role in the formation of the ground thermal regime of this site.

- During the snow-free period the main factor that influences the ground surface thermal regime is the water on the surface and in the near-surface soils (high water content decreases the ground surface temperatures due to the heat loss through evaporation).

- The complete freeze-up of the AL was observed every year for the whole period of observations at the SL-1, SL-4 and the SL-2 sites. At the SL-3 site, the ground thermal regime turned out to be strongly dependent on the snow depths and air
temperatures. So, in some years with warmer air temperatures and thick snow cover, the AL at this site did not freeze-up completely. During these years, a shallow “talik” was developed above the permafrost surface. During the colder years or the years with shallower snow, this “talik” disappeared.

- Existence of the extremely cold ground temperatures (that were observed at the SL-4 site during the entire period of measurements) may enable the formation and preservation of the ice wedges, which are considered to be not typical for this region.
- The thermal offset turned out to be the highest at the SL-3 site (-2.8°C) because of the highest heat turnovers due to the high water content in near-surface soils and hence big differences in the thermal conductivity of soils in the thawed and frozen states.
CHAPTER 6

Conclusions

Spatial variability of the climate and permafrost conditions in the Fairbanks region was studied using temperature measurements at a number of sites in an area of approximately 100x100km.

The effect of climatic and surface parameters (air temperature, vegetation and snow cover, soil properties including water content) on the thermal regime of the ground was studied using the temperature and water content measurements at five sites that form the Smith Lake-profile. These sites represent different landscapes and hence various microclimate and permafrost conditions.

The effect of the topographic gradient and the slope aspect on the thermal regime of the ground was studied using the temperature measurements and site moisture characteristics at three local topo-sequences in the Fairbanks region.

The following conclusions can be made based on the analysis of the obtained results:

- The only observed pattern in the distribution of the air, ground surface and ground temperatures in the Fairbanks region were observed along the local topo-sequences. However, this conclusion could change when more spatially distributed data will become available;

- Strong air temperature inversions during the cold period are present in the Fairbanks region and are strongly dependent on the topographic gradient (position on the slope) and not on the absolute elevation of the topo-sequence;
The main factors that determine the ground surface and ground thermal regime in the Fairbanks region during the cold period are the air temperature, snow cover, the combined effect of the snow cover and the moisture content at the surface and in the near-surface soils, and the ground vegetation;

Direct effect of ground vegetation was observed in the case of the Potato Field site, where the absence of the moss cover results in the absence of permafrost, at least in the upper few meters;

Vegetation cover and microrelief features (tussocks in case of the Smith Lake-4) also influence the ground surface and ground thermal regime indirectly through the distribution of the snow cover and ability to retain moisture;

Some local events like melting of the first snow can cause drastic changes in the ground thermal regime;

The ground surface temperature and ground thermal regime during the snow-free-period are determined mainly by the air temperature and water content at the surface and in the near-surface soils due to the effect of evaporation;

Variability in the ground surface and ground temperatures can be higher within each site due to the local microrelief features and snow distribution than along the whole topo-sequence;

Slope aspect and topographic gradient influence the ground surface and ground thermal regime through the snow cover distribution and the air inversions during the cold period and through the water content at the surface and in the near-surface soils during the warm period;
➢ The thermal offset is mostly dependent on the water content in soils because of the high heat turnovers due to significant differences in the thermal conductivities in the thawed and frozen state;

➢ The snow surface temperature is very close to that of the air, hence in some calculations assumption that they are equal can be substantial.
References


