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MASTERS THESIS

# Modelling permafrost temperature response to variations in meteorological data

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Cover photo (source : F. Caline) : Construction in the high Arctic: tent-camp on an ice-cored moraine during field session (Drønbreen, Svalbard, Norway)

This thesis was done from March to August 2000 in Longyearbyen, Svalbard. It is submitted to the Ecole nationale des ponts et chaussées (Marne-la-Vallée, France) for the degree of graduate engineer and to the DEA Sciences de l'information géographique (Marne-la-Vallée, France) for the degree of Ms. Sc.
The thesis combined geography, for the litterature work, and geotechnical science, for both the litterature work and the modelling.
The thesis is part of the SIP7 programme (1999-2003) of the Norwegian geotechnical institute (Oslo, Norway). The main objective of the SIP7 programme is to investigate how the permafrost responds to different loads such as terrestrial pollution and industrial activity and to establish reliable, effective and environmentally safe solutions for construction on permafrost and remediation of polluted areas.
The SIP7 programme is linked to the PACE (Permafrost and Climate in Europe) project (1997-2000), funded by the European union environment and climate research programme and the Swiss government. For this project, 100 m deep holes were drilled in permafrost areas of Europe, including one situated near Longyearbyen, Svalbard.
This thesis was presented during the International Permafrost Workshop, june 2000, in Longyearbyen.
During fall 2000, an article based on the results from this thesis will be written (Instanes and Caline) for the Norwegian Journal of Geography.

Longyearbyen, Svalbard, Friday the 1st of September, 2000

Fabrice Caline

## Abstract

In 1977, the Norwegian Geotechnical Institute established a permafrost research station at Sveagruva, Svalbard (77º 54'N, 16º 41' E). Meteorological data and ground temperatures have been recorded continuously since the start in 1977. Measurements are carried out automatically with data recordings every hour. Measurements are taken of air temperature, wind force and wind direction, ground temperature down to 8 m below the soil surface, incoming and outgoing radiation, global radiation, heat flux in the ground and humidity of the air.

The main objective of the thesis was to gain a better understanding of the physical processes determining permafrost temperatures and possible short-term and long-term changes in temperature at depth.

From a geotechnical point of view, this is important for the design of infrastructures on permafrost, since a possible global air temperature increase may affect the stability of existing structures and the design of future structures.

From a geographical point of view, a change of thermal regime of the permafrost has direct consequences on the landscape.

A finite-elements software was used for the modelling of the ground response to a temperature change. The model was calibrated with temperature series recorded at Sveagruva but the results were not considered precise enough to draw any conclusion on the effect of a global air temperature increase on the behaviour of the permafrost.

However a simulation was run to model a 2 ºC increase of air temperature on a 50 years period. Based on the results from this simulation, the active layer was estimated to deepen by 25 cm.

## Preface

In the two first chapters of this thesis, most scientific material was found in the books of Phukan (1985), of Andersland and Ladanyi (1994) and of Washburn (1979).

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# Chapter 1 – Frozen ground and permafrost

## 1.1 Seasonally frozen ground

Seasonally frozen ground "...is ground frozen by low seasonal temperatures and remaining frozen only through the winter" (Muller, 1947). This definition includes the ground layer overlying permafrost that undergoes an annual cycle of freezing and thawing, known as the active layer. However, the term is generally meant as restricted to a non-permafrost environment, characterising areas with frost penetration of at least 30 cm once in 10 years (Koster, 1993).

The area of occurence of frozen ground (seasonally and perennially) has been estimated to cover 48 per cent of the total land area in the Northern Hemisphere, its southern limit being near lat. 40 degrees (Bates and Bilello, 1966). Of this area, 46 per cent are assumed to be covered by permafrost (i.e. perennially frozen ground), exclusive of mountains (Stearns, 1966, 9-10) and thus 54 per cent by seasonally frozen ground.

The depth of seasonal freezing is increasing from a few millimeters in southern regions to 3 m in Canada (Crawford and Johnston, 1971). Measuring such depths is not always practicable and therefore several indirect approaches have been suggested. One possibility is to use the Stefan equation, which, according to Aldrich and Paynter (1966), may be written:

$$z = \sqrt{\frac{2 \cdot k_f \cdot I_{sf}}{L}} \quad \text{equation 1-1}$$

where $k_f$ is the frozen soil thermal conductivity (kJ · day$^{-1}$ · m$^{-1}$ · ºC$^{-1}$), $I_{sf}$ is the surface freezing index (ºC · day) and $L$ is the latent heat of the soil (kJ · m$^{-3}$). Methods for calculating the surface freezing index $I_{sf}$ are detailed in Chapter 2.

Thermal and mechanical properties of soils are radically affected by ground freezing. Permeability, for example, is decreased, resulting in increased run-off, with vast implications for flooding and water supply.

## 1.2 Permafrost

### 1.2.1 Overview

The term permafrost, also known as pergelisol (Bryan, 1946) and perennially frozen ground, was first defined by S. W. Muller (1947) :

"Permanently frozen ground or permafrost is defined as a thickness of soil or other superficial deposit, or even of bedrock, at a variable depth beneath the surface of the earth in which a temperature below freezing has existed continually for a long time (from two to tens of thousands of years). Permanently frozen ground is defined exclusively on the basis of temperature, irrespective of texture, degree of induration, water content, or lithologic character."

The term has been widely adopted but defined as perennially rather than permanently frozen ground, since changes in climate and surface conditions can cause rapid thawing of permafrost.

It is important to note that this definition uses 0 degrees as the basic criterion, while in practice the soil freezes at lower temperatures, depending mainly on the amount of salt in the pore water.

### 1.2.2 Active layer

In a permafrost environment, the top layer of ground undergoes an annual cycle of freezing and thawing similar to that of seasonally frozen ground. This top layer is defined as the active layer and its thickness varies from as little as 15 cm in the far north to several meters to the south (ca. 90 cm in Svea), depending on many factors including mean annual air temperature (MAAT), soil type and surface cover (snow, vegetation).

The active layer is responsible for ground heave when it freezes in the winter. Under this freeze process, the pore water will increase in volume by about 9 per cent, and as most soils are inhomogeneous, the heave process will not be uniform along the surface. This, of course, strongly affect constructions, especially highways, which usually experience an increased surface roughness and bumps. The most important damages to the pavement are, however, caused in the thawing period when excess water cannot drain downward through still-frozen soil and, hence, pore-water pressure is high.

### 1.2.3 Permafrost geographical distribution

Permafrost generally occurs in places where day air temperatures are below 0 ºC for three fourths of the year, below –10 ºC for half of the year, and rarely over 20 ºC. Precipitations are characteristically below 100 mm in winter and below 300 mm in summer.

The permafrost region of the world is subdivided into a continuous zone, where permafrost occurs everywhere, and a discontinuous zone, where permafrost occurs only in some areas (Figure 1-1). In the southern parts of the discontinuous zone, permafrost becomes thinner and breaks up into patches as it merges with seasonally frozen ground where permafrost is lacking.

The thickness of permafrost is about 60 to 90 m at the southern limit of the continuous zone and increasing to over 1000 m in North-East Siberia (Table 1-1).

The limit between the continuous and the discontinuous zones is arbitrary set to the -5 ºC isotherm measured at a depth of zero or negligible temperature fluctuation (called the level of zero annual amplitude).

| Location | Depth (in meters) |
|---|---|
| Alaska Fairbanks | 30 to 120 |
| Prudhoe Bay | 610 |
| Canada Mackenzie Delta | 90 |
| Resolute, Cornwallis I. | 935 |
| Greenland Thule | 520 |
| Svalbard Longyearbyen | 190 |
| Sveagruva | 125 |
| Russia Yakutsk | 200 to 250 |
| Noril'sk | 325 |
| Markha River, upper reaches | 1450 to 1500 |

Table 1-1 Permafrost depth in the Northern Hemisphere

source: Washburn, 1979, table 3.5; Gregensen and Eidsmoen, 1988 (for data on Svalbard)

Considering offshore (also known as subsea) permafrost, we can actually map a third permafrost zone. That type of permafrost was observed by Werenskiold on Svalbard in the beginning of the century (1922) and, more recently, investigation have proven the existence of subsea permafrost in Alaska, Canada and Northern Russia. The thickness of subsea permafrost may vary widely as its existence is directly related to the history of the different shelves : much of the subsea permafrost probably originated on land and became submerged by postglacial rise of sea level and coastal erosion.
Subsea permafrost causes several problems for oil exploitation and has therefore been studied more in detail recently.


Figure 1-1 Permafrost areas in the Northern Hemisphere (source: www.AMAP.no)

According to Haeberli et al. (1993), permafrost is also estimated to occupy some 5 million $km^2$ in mountain ranges at middle and low latitudes, excluding the uplands of Eastern Siberia and the Far East.

In the Antarctic, permafrost is limited to areas where the earth's surface is covered by seasonal or perennial snow cover and by a "cold" ice sheet having a temperature below 0 ºC at the base (Baranov, 1959). The presence of subsea permafrost is high in these regions as, in McMurdo Sound for example, water temperatures are about -1.8 ºC.

### *1.3 Periglacial forms and frost-action processes*

Permafrost regions are characterized by a number of physical features (ice wedges, pingos, thermokarst terrain) which are indicators of ground ice.

Ground ice is defined as ice present in permafrost, regardless of amount or configuration. The voids in permafrost are generally filled with ground ice. In fine-grained soils, excess ice, called segregated ice, is also formed by moisture migration under different thermal gradients. Most large masses of segregated ice are horizontal layers found more than 3 m under the surface (Mackay and Stager, 1966), the maximum observed depth being of at least 60 m (Mackay, 1976).

Understanding of ground ice conditions and their induced surface features is important from the engineering point of view. It gives information on the physical and mechanical properties of the ground while excess water, released by segragated ice when the ground is thawed, has to be taken into account when designing and constructing structures in frozen ground. Such structures are indeed likely to cause thaw settlement or modify soil conditions.

The terrain features associated with ground ice typically affect the geomorphology of permafrost regions. The most important features are ice wedges, pingos and thermokarst terrain.

#### 1.3.1 Ice-wedges

An ice-wedge (Figure 1-2) is initiated by small shrinkage cracks which occur at the ground surface when air temperature falls down in the winter. The cracks result from soil contraction and their size depends on the coefficient of thermal contraction $a$ and the rate and magnitude of air temperature change. They are usually only a few millimeters wide but may extend downward several meters. In the spring, snow meltwater fills the cracks, freezes, and forms a vertical ice vein that penetrates the permafrost. The following winter, the cracks widen under new ground contraction. This cycle is repeated over several hundred years.

Ice wedges often create a network of polygons on the ground surface similar to the surface pattern formed by cracks in drying mud. Such ice-wedge polygons are one of the most common forms of patterned ground. Other patterned ground forms include circles and strips.

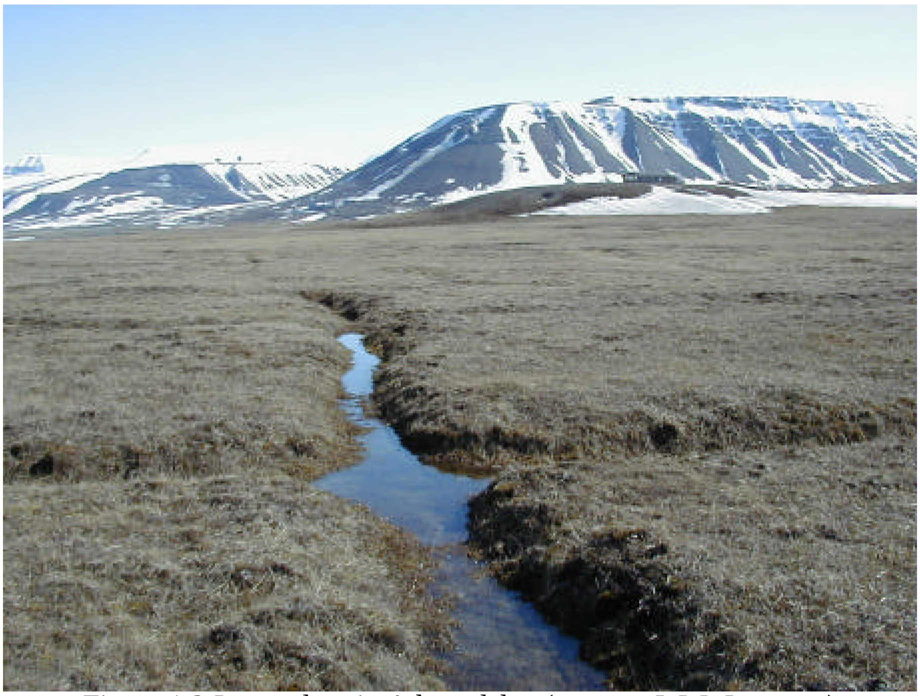


Figure 1-2 Ice-wedges in Adventdalen (source : J. W. Jeppesen)

### 1.3.2 Pingos

A pingo is a conical, more or less asymmetrical mound or hill over 10 m high, with a circular or oval base over 100 m across and commonly fissured at the summit. Pingos can be up to 50 m high and 150 m across and consist of a soil cover overlaying a massive ground ice core. It is interesting to note that pingos have long been used as sites for geodetic triangulation stations.

A closed-system pingo builds up when permafrost has formed over an entire lake basin. As more water freezes and finds its expansion room limited by impervious frozen material, a considerable uplift pressure is created, raising a blister.

The open-system pingo is more common in hilly terrain, as on Greenland or Svalbard for example. Water is supplied by springs where artesian pressure has developed in unfrozen permafrost zones. This water penetrates the ground under the action of the hydraulic gradient resulting from the differences in elevation. Due to the impermeability of the permafrost, the water stays at the surface where it freezes. The continuous water supply allows buildup of a considerable ice mass, which domes the ground surface upward over a period of many years.

### 1.3.3 Thermokarst

Thermokarst is a generic term for terrain features resulting from the differential melting of ground ice. These features include mounds, caverns, disappearing streams and large flat-floored valleys with steep sides. Thermokarst are due to climatic changes, either local, for example induced by changes in vegetation (farming, constructions, fire), or more general such as climatic change.

In Fairbanks, Alaska, the clearing of trees and vegetation in the 1920s led to the development of thermokarst mounds varying from 3 to 15 m in diameter and 0.3 to 2.4 m in height (Rockie, 1942). These mounds resulted from the thawing of the ice-wedged polygons and the induced soil collapse.

Thermokarst resulting from climatic change is more likely to be associated with the discontinuous zone where the thermal balance of permafrost is more delicate, with changes being apparent within 20 years (Thie, 1974).

# Chapter 2 – Heat flows and temperature in permafrost

## 2.1 Heat transfer theory

### 2.1.1 Fourier's law of heat flow

The principal mechanism for heat flow in soils is conduction. Conduction is the mechanism of heat transfer, through soil particles or soil pore fluids, from a warm part of the mass to a cooler part. The rate at which heat is transferred by conduction is governed by Fourier's law

$$\underline{J_q} = -k \cdot \underline{\nabla T} \quad \text{(Equation 2-1)}$$

where $\underline{J_q}$ is the heat flux vector ($W \cdot m^{-2}$), $k$ the thermal conductivity ($W \cdot m^{-1} \cdot K^{-1}$), and $\underline{\tilde{N}\, T}$ the thermal gradient ($K \cdot m^{-1}$).

This equation states that heat flow due to conduction is directly dependent on the thermal conductivity of the soil medium and temperature gradient. The negative sign indicates that the heat flows from high temperatures to low temperatures.

Another mechanism for heat transfer in soils is convection. Hinkel and Outcalt (1994) have demonstrated that convective transfer occurs within the active layer during snowmelt and during infiltration of precipitation in summer (Hinkel et al., 1997). Convection can also occur when air masses penetrate in the upper layer of highly porous soils, either in a natural environment, for example in a rocky slope, or in engineering constructions like road gravel fills.

However in most cases convective effects are masked by conduction.

### 2.1.2 Energy balance and heat equation

The heat transfered through the soil following Fourier's law contributes to the warming or cooling of the ground. This temperature variation of the ground is governed by a behaviour law of energy conservation

$$\frac{\partial}{\partial x_i}\left( k_{x_i} \frac{\partial T}{\partial x_i} \right) + Q = l \, \frac{\partial T}{\partial t} \quad \text{(Equation 2-2)}$$

where $T$ is the temperature (K), $\underline{k}$ is the vector of thermal conductivity ($W \cdot m^{-1} \cdot K^{-1}$), $Q$ is the applied boundary flux ($W \cdot m^{-3}$) and *?* is the capacity for heat storage of the ground ($W \cdot m^{-3}$).

This equation states that the difference between the heat flux entering and leaving an elemental volume of soil at a point in time is equal to the change in the stored heat energy. The stored heat energy governs, in turn, the variation of temperature of the elemental volume of soil through the ground thermal properties (capacity for heat storage *?*).

The capacity of a soil to store heat is composed of two parts : the volumetric heat capacity of the material and the latent heat associated with the phase change

$$l = c + L \cdot w_v \cdot \frac{\partial w_u}{\partial T} \quad \text{(Equation 2-3)}$$

where $c$ is the volumetric heat capacity ($W \cdot m^{-3}$), $L$ is the latent heat of water ($3.34 \cdot 10^8$ $J \cdot m^{-3}$), $w_v$ is the volumetric water content of the soil, $w_u$ is the unfrozen water content ($0 = w_u = 1$) and $T$ is the temperature (K).

The volumetric water content of the soil, $w_v$ , is defined as the ratio of the volume of water to the total volume of soil and the unfrozen water content, $w_u$ , is the percentage of the water that is unfrozen. Soil properties are explained more in Chapter 4.
Substituting for ? in Equation 2-2 leads to the complete differential equation

$$\frac{\partial}{\partial x_i}\left(k_{x_i} \cdot \frac{\partial T}{\partial x_i}\right)+Q=\left(c+L \cdot w_v \frac{\partial w_u}{\partial T}\right)\frac{\partial T}{\partial t} \quad \text{(Equation 2-4)}$$

### 2.1.3 Boundary conditions

Apart from air temperature, the major source of heat transfer at the ground surface when no convective heat is involved, is solar radiation (measured in $J \cdot m^{-2}$). The amount of solar radiation absorbed by the ground surface depends on the absorptivity of the surface, which ranges from 0.20 for fresh snow to 0.60-0.70 for concrete, 0.80-0.90 for a wet soil surface and 0.90-0.95 for a forest (Scott, 1964).
As is discussed more in detail in Chapter 3, the influence of meteorological factors on the heat flux at the surface of the active layer is however dampened by the surface cover : snow and vegetation.
Below the depth of zero annual amplitude, the heat flux is constant and proportional to the norm of the geothermal gradient $i_g$ ($K \cdot m^{-1}$). In Sveagruva, considering a geothermal gradient of 0.05 $K \cdot m^{-1}$ (Gregensen and Eidsmoen, 1988) and a conductivity of the frozen ground of 1,16 $W \cdot m^{-1} \cdot K^{-1}$ (Berggren, 1983), the heat flux is therefore

$$\left|\underline{J_q}\right| = -k\left|\underline{i_g}\right| = -1.16 \cdot (-0.05) = 0.058 \ W \cdot m^{-2} \quad \text{(Equation 2-5)}$$

## *2.2 Temperature profile in permafrost*

The variations of the air temperature throughout the year generates fluctuations in the soil temperature. The soil and its possible cover (snow, vegetation) behave like a buffer and the amplitude of the temperature fluctuations will therefore decrease with depth down to a point of zero amplitude, defined as the depth at which the ground temperature fluctuates by less than 0.1 ºC per year. The temperature at this depth is called mean annual ground temperature (MAGT) and is usually up to several degrees warmer than the mean annual air temperature (MAAT). When plotted against depth in the ground, the monthly average temperatures give a characteristic curve called *whiplash curve* because of the way the monthly averages lash to and fro as the seasons change.
The depth of zero annual amplitude is commonly between 10 and 15 m and the temperature of this point ranges from – 1 ºC to about – 8 ºC.
Below this depth the geothermal gradient ranges from 1 ºC per 22 m (0.045 $K \cdot m^{-1}$) to 1 ºC per 160 m (0.006 $K \cdot m^{-1}$) with an average value of 1 ºC per 30 m. Measures made by Gregensen and Eidsmoen (1988) give a geothermal gradient of 1 ºC per 30 m in Longyearbyen and 1 ºC per 20 m in Sveagruva (Svalbard).
The term steep gradient commonly refers to a high $K \cdot m^{-1}$ ratio.
The geothermal gradient is inversely proportional to conductivity which, in turn, has a strong influence on permafrost thickness. Observations made in Alaska at three places with similar mean annual (ground) surface temperature (MAST, measured 1 cm below the soil surface) reflect this behaviour very obviously: the permafrost at Prudhoe Bay is 50 per cent thicker than at Barrow and 100 per cent thicker than at Cape Simpson.

The thickness of stable permafrost, i.e. that is neither aggrading (building up) or degrading (thawing), can be calculated with the Terzaghi formula (1952) :

$$H_p = \frac{MAST}{-i_g} \quad \text{(Equation 2-6)}$$

where $H_p$ is the thickness of the permafrost (m), *MAST* is the mean annual (ground) surface temperature (ºC) and $i_g$ is the norm of the geothermal gradient ($K \cdot m^{-1}$).

### *2.3 Active layer thermal regime*

The limit between the permafrost and the active layer, called permafrost table, has its depth controlled by the thermal regime of the active layer. The permafrost table is a rigid surface capable of bearing considerable loads without deforming and is, as such, of great importance for the construction of structures. It is also an impermeable layer, which gives way to a high moisture content and the formation of ice lenses in the overlying active layer.

The zone immediately above the permafrost table, which is situated at the limit of summer thawing, experiences a considerably long thawing period (as long as 115 days a year) caused by the latent heat of fusion of ice ($3.34 \cdot 10^6$ $J \cdot m^{-3}$). This phenomenon is called the zero curtain and is a notable aspect of the thermal regime of the active layer.

The freezing of the active layer is actually a dual cooling process from above, through meteorological and geographical factors (air temperature, snow and vegetation cover, solar radiation), and from below through the permafrost table (upfreezing). External factors are discussed more in detail in Chapter 3. Upfreezing is, according to Mackay's observations (1973) in the Mackenzie Delta area of northern Canada, the greatest where the active layer is thickest. The amount of upfreezing found by Mackay ranged from 2 to 13 cm.

The structure of the ground has also a strong influence on the thermal regime of the ground. The conductivity of ice is about 4 to 5 times that of the soil and ice wedges will therefore accelerate the rate of heat transfer into the active layer during the thawing period. Massive ice beds, on another hand, tend to delay the thawing of the ground due to the latent heat of fusion of ice.

Inversely, the structure of frozen ground is affected by its thermal regime. There are, for example, generally smaller ice masses in the active layer than in permafrost.

In some places where the ground is an old seabed, which is the case in Sveagruva, the salinity of the water is approximately that of seawater, i. e. 34 $g \cdot l^{-1}$. This modifies considerably the temperature of fusion of the ice, which can be estimated using an empirical equation developed by Velli and Grishin (1983). According to these authors the temperature shift, *?T*, due to salinity, can be determined from

$$\Delta T = T_k \cdot \left( \frac{S_n}{1000 + S_n} \right) \quad \text{(Equation 2-7)}$$

where $S_n$ is the salinity ($g \cdot l^{-1}$) and $T_k$ is a reference temperature equal to 57 K for sea salt, 62 K for NaCl and 32.5 K for $CaCl_2$.

For $S_n$ equal to 34 $g \cdot l^{-1}$, as in Sveagruva, equation 2-7 gives *?T* equal to 1.9 K, meaning the temperature of fusion of the ice is of –1.9 ºC.

Even though the soil is defined as frozen when it has a temperature below 0 ºC, its mechanical properties are directly dependent on its salinity : increased salinity reduces the ice content, thereby reducing the frozen soil strength and increasing its creep rate at a given temperature.

# Chapter 3 – Permafrost response to changing climate

Of all factors that affect the evolution of the permafrost, air temperature is the fundamental one. In the context of a global warming of the Earth, it is therefore important to analyze the consequences on the permafrost of a climate change.

Contrary to the precedent chapter, where the focus was put on the thermal regimes of the active layer and of the permafrost under known boundary conditions, in this chapter we will analyze the climatic processes that take place above the ground surface (in particular the role of the snow and vegetation) in order to assess the consequences of a climate change on the thermal regime of the ground. Further, we will discuss the response of the ground, in terms of mechanical properties, to a modification of its thermal regime. At last, we will present results from recent researches that have interpreted temperature profiles in continuous permafrost in terms of indicators of climate change.

## *3.1 Interaction of climate and ground surface temperature*

The ground surface temperature (measured at 1 cm depth) is a very convenient boundary condition when calculating the ground temperature profile, for example with a finite-element method. However the most accessible temperature data are not surface temperatures, rather air temperatures. The problem is that ground surface temperature is not only influenced by air temperature but also by other factors like solar radiation and snow and vegetation cover. As a result, mean annual ground surface temperatures differ from mean annual air temperatures with no constant difference between them (Brown, 1963), and wide variations in ground thermal conditions can occur within small areas of uniform climate. Predicting effects of a climate warming on permafrost begins therefore with understanding the processes that take place on top of the active layer, in particular the role of snow and vegetation.

### 3.1.1 Impact of snow and vegetation cover

The snow cover plays an important role in heat transfer to the ground surface, due to its high degree of insulation. Depending on how compact the snow is, its conductivity ranges from 5 per cent, for dry new snow, to 25 per cent, for old compacted snow, of that of wet sand (Gold and Lachenbruch, 1973).

Calculations made by Goodrich (1982) showed that a doubling of snow cover from 25 to 50 cm increased the minimum ground surface tmperature by about 7 ºC, and the mean annual surface temperature by 3.5 ºC.

Other studies, made near Schefferville, Labrador, Canada, assessed that snow cover variations are responsible of 70 percent of the variance of ground surface temperatures (Nicholson and Granberg, 1973).

As a rule, ground surface temperature increases with snow cover thickness, due to insulation from air cooling. However, over a certain thickness, the effect of the snow cover is inversed as it will take longer to melt and therefore less warming will occur in summer. This remarkable behaviour is illustrated by measurements of King (1983 1986) in the Tarfala region in northern Sweden (MAAT –3.5 to –4.0 ºC). Permafrost was regularly found at sites where snow cover in winter was les than 1.1 m, was missing at sites where snow cover thickness reached values between 1.4 and 2.4 m, but occurred again at sites where the thickness increased to more than 2.4 m.

Vegetation plays also an important role in permafrost areas, mainly by shielding the ground surface from solar heating in the summer. This has been very well illustrated in an experience made by Linell (1973) in an area of discontinuous permafrost near

Fairbanks, Alaska. After all trees and brush were removed a rapid increase in depth of the permafrost table was observed, proving clearly the influence of the vegetation cover.
It is however important to note that, in polar areas of continuous permafrost, the role of the vegetation cover is limited since vegetation in such areas is limited if not almost unexistent.

### 3.1.2 Modelling the role of the surface cover – freeze/thaw indexes and *n*-factors

The most widespread way of modelling the complex dependence between air temperature and ground surface temperature is to use the degree-day concept. Two indexes are defined based on air temperature: the air thawing index, $I_{at}$, and the air freezing index, $I_{af}$. Over one season, the air thawing index is the number of degree-days of positive air temperature while the air freezing index is the number of degree-days of air temperature below 0 ºC. Both indexes are calculated from daily average air temperatures. For example, a mean daily air temperature of +5 degrees for 3 days gives an air thawing indes equal to +15 ºC · days.
Similarly, surface indexes are defined: the surface thawing index, $I_{st}$, and the surface freezing index, $I_{sf}$. These surface indexes can be calculated through the thaw and freeze n-factors, empirically defined respectively as the ratio of the ground surface thawing index ($I_{st}$) to the air thawing index ($I_{at}$) and as the ratio of the ground surface freezing index ($I_{sf}$) to the air freezing index ($I_{af}$) :

$$n_t = \frac{I_{st}}{I_{at}} \qquad n_f = \frac{I_{sf}}{I_{af}} \qquad \text{(Equation 3-1)}$$

Approximate *n*-factors for several surface types are listed in Table 3-1. For such surfaces, surface indexes can be estimated from the combination of the corresponding *n*-factor and air index. Surface indexes are for example used to calculate the depth of seasonal freezing/thawing, as detailed in Chapter 1.
The *n*-factors are of course directly dependent on vegetation cover.

| Surface | Freezing, $n_f$ | Thawing, $n_t$ |
|---|---:|---:|
| Snow | 1.0 | - |
| Pavement | 0.9 | - |
| Sand and gravel | 0.9 | 2.0 |
| Turf | 0.5 | 1.0 |
| Spruce trees, brush, moss over peat soil | 0.29 (under snow) | 0.37 |

Table 3-1 some examples of *n*-factors. (After Andersland and Ladanyi, 1994)

## *3.2 Expected climate change and its effect on ground thermal regime*

Models of climate change generally predict a warming of the air temperature and an increase in precipitations. In the studies we refer to in this paragraph, different scenarios of climate warming have been considered. However, as those scenarios are only suppositions, the results they lead to should be considered qualitatively more

than quantitatively. Moreover our goal here is limited to assessing the effect of a possible climate change on the development of the permafrost, not to discuss the probability of occurence of the different scenarios of climate change that we consider for our analysis.
As discussed in the precedent paragraph, the relation between air temperature and ground thermal regime is complex and, most of all, highly dependent on local factors. A first approach is therefore to begin with a general simple model of the ground thermal regime that does not integrate the role of the surface cover nor the thawing and freezing processes in the active layer. Such a model was implemented by Lunardini (Lunardini, 1996), using the scenario for temperature change from the Intergovernmental panel on climate change (IPCC 1990) : temperature increase of 2.5 ºC in the next 55 years and of 5.0 ºC in the next 100 years. The results from such a model, adjusted with local factors, show a fairly small probability of significant permafrost disappearance during the next 55 years in Alaska. Most exposed is relict permafrost, which, as defined by Lunardini, is "permafrost that exists in regions where the present earth/atmosphere energy balance is not favourable for permafrost". The thickness of such permafrost melted completely in 55 years would most likely range from 3 to 13 m. On a 185 year basis Lunardini's model gave essentially the same results as those found by Osterkamp (1983).
The thawing process of the permafrost is detailed by Koster (1993). First the active layer increases. Second the temperature profile in the permafrost will adjust itself to the new mean annual ground surface temperature (MAST). Response times of the active layer, depending on local factors, are in the order of years to tens of years. At the permafrost table, response times are in the order of tens to hundreds of years while at the permafrost base, they are in the order of hundreds to thousands of years.
Anisimov (1989) simulated changes in the extent of permafrost in Russia, using constant local factors. The time scale was set to 50 years, with a warming in the Arctic of 7-9 ºC in winter and 4-6 ºC in summer. Results from this model suggest a 15-20 per cent reduction in the area occupied by continuous permafrost at the end of the period. The rate of the permafrost boundary retreat was estimated to vary from 10-15 $km \cdot year^{-1}$ in western Siberia to 20-25 $km \cdot year^{-1}$ in eastern Siberia. The depth of the active layer was predicted to increase by 0.5-0.8 $m \cdot year^{-1}$ during the first years and to decrease exponentially over time.

### *3.3 Consequences of a modification in ground thermal regime*

As estimated by the models related in the precedent paragraph, we can expect that tens of thousands of square kilometers of permafrost in the discontinuous area would eventually disappear under the climatic warming predicted, although complete degradation would take many centuries. In the continuous area, ground temperatures would rise and the active layer thicken, but the ground would mainly remain frozen. However this does not mean that continuous as well as discontinuous permafrost areas will be without environmental or engineering problems. The effect of a temperature rise affect indeed both the mechanical properties of the ice and the unfrozen water content in the ground, thus modifying the mechanical properties of the ground itself.
According to Smith (1988) it seems therefore reasonable to suppose that features of slope failure and related phenomena of mass-wasting in the permafrost environment would occur as a result of changes in climate and degradation of permafrost. Such phenomena are described by French (1976) and arise as a result of thawing with consequent loss of strength of the ground. They include mudflows and slumping. Soliflucion is also prominent, the impermeability of the permafrost leading to

conditions of saturation in the active layer. Finally, skin flows, i. e. active layer glides, occur under the thawing of the ice-rich contact zone at the permafrost table. They result in the detachment and movement of a thin veneer of vegetation and mineral soil and would be promoted by surface disturbance (e.g. forest fire) or climatic change.

Other environmental consequences include the spreading of thermokarst terrain, which would change the local drainage patterns.

Engineering problems would also occur as a result of changes in the mechanical properties of the ground. In northern construction, where piles are widely used, the bearing capacity of the ground would become critical. Terrain instability would also lead to major damage on roads, railways, air strips, dams, reservoirs and pipelines in affected areas.

Finally, although the present understanding of mountain permafrost does not allow predictions with any certainty, a degradation of ice-rich permafrost could take place on steep mountain slopes with a direct impact on construction work, ski runs, avalanches, floods, debris flows and rock falls (Haeberli et al., 1993)

### *3.4 Using permafrost for monitoring climate change*

In this paragraph we consider the use of permafrost to track changing climatic conditions on a 100 year scale. modifications of the permafrost environment could also, as related by Haeberli et al. (1993), be used on a longer time scale to evaluate late-glacial temperature increase.

Lachenbruch et al. (1988) studied temperature profiles to track climate changes in the past century. This study is based on ground temperature data from several oil exploration wells in Alaska. In most wells, the temperature profile of the permafrost is composed of an undisturbed zone with a constant geothermal gradient, overlain by a zone showing a warming anomaly in the upper 100 m or so. The zone below 100 m depth is, of course, not perfectly undisturbed. However past perturbations have small effects on the geothermal gradient : the effect of the “little ice age” (temperature anomaly of –0.5 ºC between 1550 and 1850 A.D.) is evaluated to cause a gradient perturbation of 1 ºC/km (the geothermal gradient being of the order of 30 ºC/km) while gradient effects of the late Wisconsin glaciation are of comparable magnitude.

The conductivity of the permafrost is obtained directly from the geothermal gradient of the undisturbed zone. However when, as it is the case for some of the wells, such a zone does not exist, it is difficult to tell if the variations of the geothermal gradient are due to climatic change or to a change in the conductivity of the ground. For such wells, ground sampling is required before the data can be used.

The geothermal gradients are used to extrapolate the theoretical temperature profiles for an undisturbed air temperature, and the warming anomaly in the upper 100 m are then interpreted as a warming of the air temperature. In order to assess the amplitude of this warming, Lachenbruch et al. considered several air temperature increase scenarios and evaluated their effect on the ground thermal profile through a simple homogeneous one-dimensional model. Even if the results obtained vary between the different sites, it is clear that there has been a consistent warming of a few ºC at the permafrost surface during the last century in the northern part of Alaska.

The use of geothermal gradients for studying past climate changes requires a good knowledge of the properties of the ground. The main advantage of this method is that it makes possible the tracking of climate changes over a long period of time (thousands of years) while air temperature records are seldom more than 100 years old.

# Chapter 4 – Ground properties

In this chapter ground sampling procedures followed under field work in Sveagruva are described. The aim of the field work was to obtain information on the ground composition and thermal properties at the studied borehole in Sveagruva.

## 4.1 Ground sampling

The ground sampling was performed with core-drilling equipment borrowed from Store Norske Spitsbergen Kullkompani (SNSK), the Norwegian mining company on Svalbard. Figure 4-1, below, shows a picture of the drilling process. Because of its proximity to the power station, the site initially planned for drilling could not be used. Another site was chosen which, unfortunately, showed to be a gravel pad, so it was totally impossible to take any ground sample from it.

Figure 4-1 Ground sampling in Sveagruva (source : F. Caline)

When ground samples are made, particular attention should be directed to the description of ice inclusions. Frozen cores are examined and logged as they are extruded from sample tubes at the drill site. Photographs of core samples are invaluable as a permanent record. Select pieces of the core are then retained for detailed examination in a properly equipped laboratory. The procedure for sampling of fjord ice is very similar, although not requiring as heavy equipment since the depth of the ice is generally below 1 m. Several field sessions involving UNIS technology students were carried out this winter in the proximity of Svea, on the Van Mijenfjord.

For sophisticated tests like ice distribution and unfrozen water content, undisturbed frozen samples must be prepared and handled carefully to prevent disturbance to their thermal and moisture regimes. As a rule, samples are wrapped in cellophane to protect them from loss of moisture and placed in well-sealed polyethylene bags from which the air has been evacuated to prevent sublimation. They may in addition be placed in insulated boxes or even portable refrigerators for shipment to the laboratory.

Berggren describes how the Svea clay was sampled :

“Svea clay was sampled and transported in the frozen state. Permafrost temperature was about –4 ºC to –6 ºC. Storing and transportation temperatures have varied somewhat, but have not been colder than –12 ºC, and thawing has been prevented.”

Ground sampling is a technique widely used for engineering projects in permafrost areas, where detailed information on subsurface conditions is required. Drilling is generally done to a depth equivalent to the width of the structure. Testing of the ground samples obtained include determination of mechanical properties, deformation behaviour of frozen, thawing and thawed soils.

### 4.2 Description of Svea clay

Berggren describes the Svea clay in her thesis for the degree of Doctor of Engineering (Berggren, 1983) :
"[The Svea clay] is a marine permafrost clay from Sveagruva, Svalbard (Spitsbergen). Svea clay contains vertical icelenses 0.5-1.0 cm wide in a roughly parallel pattern. In addition, there are interconnected icelenses randomly oriented, up to one or two millimeter thick. The color of the clay is dark brown while the ice is transparent. The clay was sampled in the frozen state by the USA 3" Ice coring auger.
The mineral contents are chlorite, illite, albite, kaolin, quartz (10%).
After separating the coarse icelenses from the clay, the salt content of the clay was measured to be 50-70 $g \cdot l^{-1}$ and the water content decreased to about 30 %. The icelenses only contains minor amounts of salt, if any, measured to be 0.05-0.34 $g \cdot l^{-1}$."

### 4.3 Measured values for Svea clay

Since no ground sample was taken in Sveagruva during the thesis, the data used here are taken from Berggren's thesis (Berggren, 1983). These data were obtained through three different series of tests from 1980 to 1981 (Berggren, 1980, T. Johansen et. al., 1981, T. Furuberg, 1981). During this period a total of 27 core samples were prelevated from 1.1 to 4.2 m under the surface.
Mean values of basic and for our model relevant ground properties are given in Table 4-1. Average values were considered through the modelling because ground properties were not observed to be significantly correlated to depth of sampling.

| Property | Bulk density ($kg \cdot m^{-1}$) | solids density ($kg \cdot m^{-1}$) | salt content ($g \cdot l^{-1}$) | porosity (%) | saturation (%) | volumetric water content (%) |
|---|---|---|---|---|---|---|
| Mean value | 1680 | 2740 | 34.2 | 58.0 | 91.6 | 53.1 |

Table 4-1 Data for Svea clay samples (from Berggren, 1983)

Volumetric heat capacity given by Berggren includes latent heat of fusion of ice. The values considered here are therefore those :
- at –30 ºC for frozen Svea clay, where ice melting is almost inexistent
- at 0 ºC for unfrozen Svea clay, where all ice is melted.

Heat capacities of water and ice are given for comparison.

| | Volumetric heat capacity ($J \cdot m^{-3} \cdot K^{-1}$) |
|---|---|
| Svea clay (Frozen) | $2.1 \cdot 10^6$ |
| Svea clay (Unfrozen) | $3.0 \cdot 10^6$ |
| Water | $4.1 \cdot 10^6$ |
| Ice | $1.9 \cdot 10^6$ |

Table 4-2 Volumetric heat capacity of Svea clay

Heat conductivity of Svea clay is given in Figure 4-2.

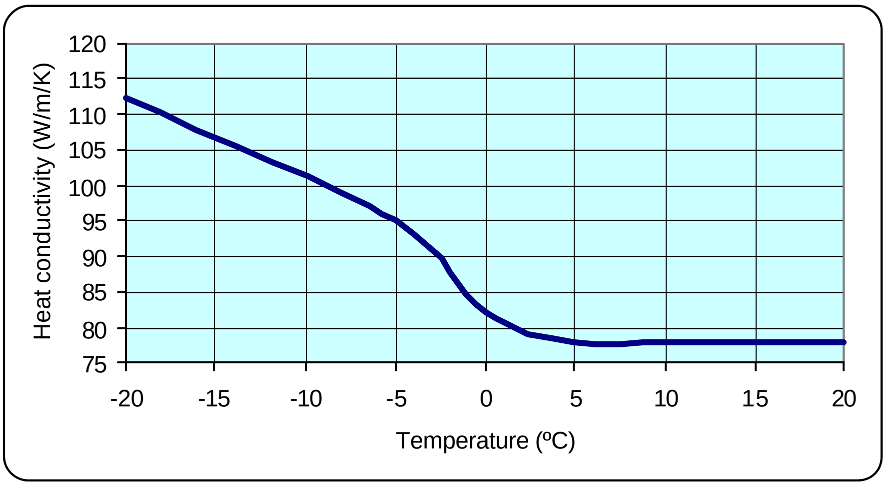



Figure 4-2 Heat conductivity of Svea clay

The unfrozen water content of Svea clay between –10 and 0 ºC is given in Figure 4-3. Below –10 ºC, the unfrozen water content is decreasing slowly with decreasing temperature, at a rate of 0.15 % $\cdot$ $ºC^{-1}$.

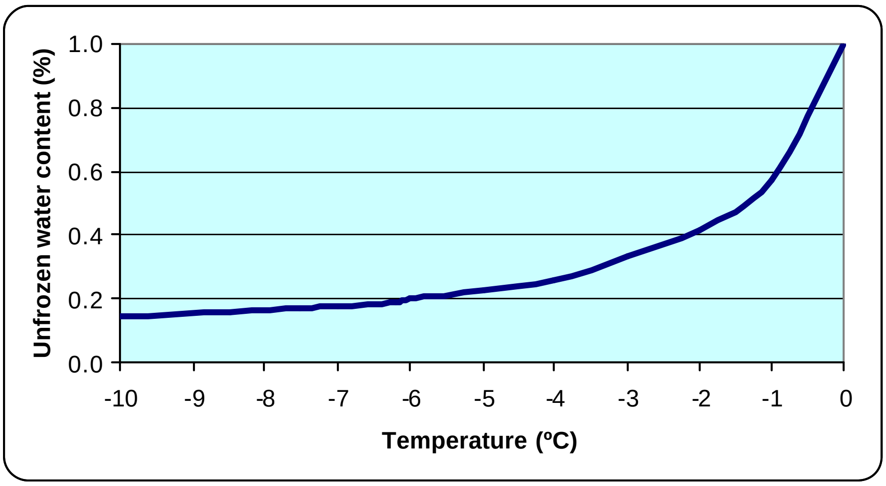



Figure 4-3 Unfrozen water content of Svea clay

## Chapter 5 – Data analysis

### 5.1 Sveagruva weather station

In 1977 the Norwegian committee on permafrost started the installation of a weather station in Sveagruva, Svalbard (Figure 5-1) in order to obtain information on the climatic and ground conditions on Svalbard (Bakkehøi and Bandis, 1988).
The station was completed in June 1978 (Figure 5-2). It is situated at 77º 54'N, 16º 41' E, 9 m a.s.l. (Figure 5-1). At this latitude, the midnight sun period lasts from 21. April to 21. August and the dark period from 27. October to 13. February. High mountains (700-950 m a.s.l.) are situated in the sector west to north. Between the station and the mountains is a plain of level tundra. Terrain southwest of the station consists of moraines, and west-southwest is the direction out Van Mijenfjorden.

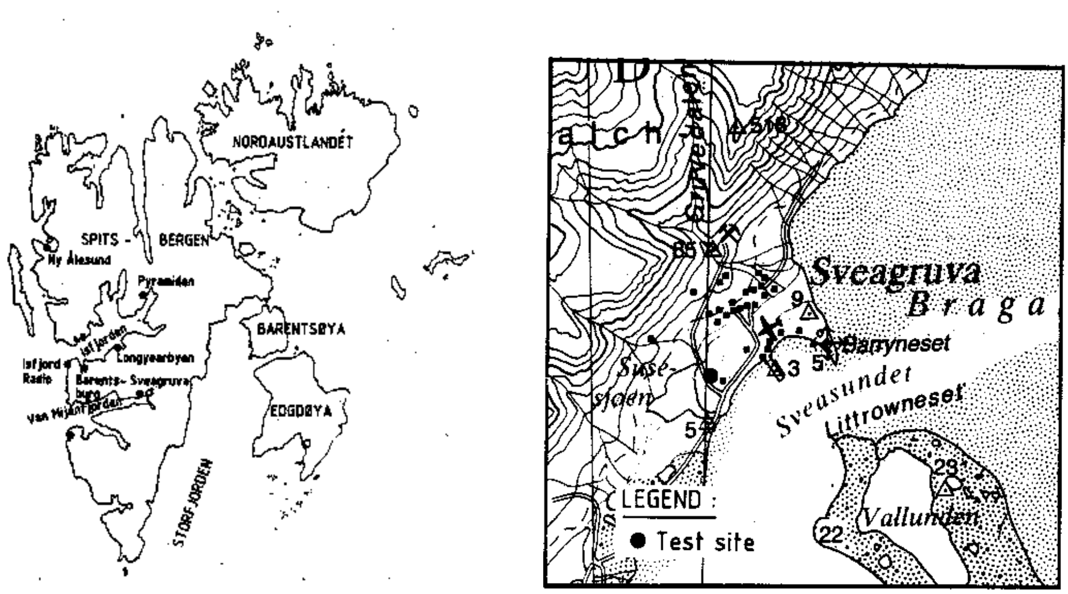



Figure 5-1 Maps of Svalbard and of the Sveagruva area

Norwegian geotechnical institute (NGI) is responsible for the maintenance of the station. Data is collected automatically through three separate Aanderaa dataloggers on an hourly basis (Figure 5-2)
Different data are measured : air temperature, wind force and direction, ground temperature at several depths below the soil surface (0, 5, 20, 50, 100, 150, 200, 250, 300 and 800 cm), incoming and outgoing radiation, global radiation, heat flux in the ground and humidity of the air.
In addition temperature is measured in the ground down to 2 m below a concrete surface of 2 m x 2 m and a thickness of 5 cm.
Finally, daily manual observations are made of the snow depth and density (Norwegian meteorological institute).

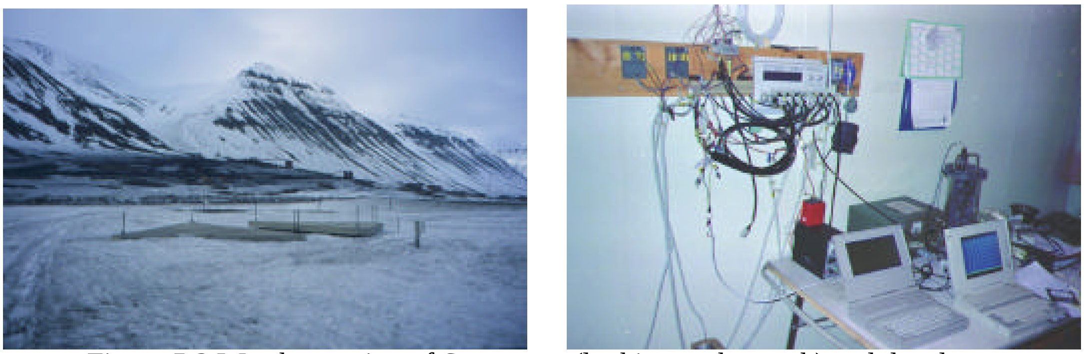


Figure 5-2 Weather station of Sveagruva (looking to the north) and data loggers

Other observations are carried out around the NGI weather station.
First, Norwegian meteorological institute (DNMI) opened a station in May 1978 and is running it with the Norwegian committee on permafrost (Fjørland et. al., 1997).
NGI is also measuring in situ long term creep rate of the ground, using six instrumented footings installed to depths varying between 1.2 and 2.2 m (Lunne and Eidsmoen, 1988).
Finally, NGI drilled a 100 m deep drillhole and installed permanent thermistors. This drillhole is part of a project which has studied permafrost conditions in the shore area of Svalbard. Two sites were chosen, Longyearbyen and Sveagruva, and at each site, two 100 m deep holes were drilled, one at the shore line, the other some distance from the shore line (resp. 500 m and 200 m, i. e. at the NGI weather station) (Gregersen and Eidsmoen, 1988).

### 5.2 Data processing and analysis

The data from the NGI weather station in Sveagruva were stored during allmost 25 years but apart from Bakkehøi and Bandis (1988), nobody had really been analysing them. As a result, they were largely inhomogeneous and a considerable processing was necessar before they could be used in a permafrost model.
The first data inhomogeneity was in the storage itself : for a given time period, there were up to four files containing measures of the same parameter (generally two files). The problem is that the values did not correspond from one file to the other. Moreover, in each file, the data contained many wrong values. The first task was therefore to sort the data from each file, removing wrong values. Then data from the same series were grouped in year files. Finally the year files were compared and the seemingly most correct data were finally kept for use in the permafrost model.
Several types of wrong values are found in the data files. First, some standard wrong values were obtained with the Fortran program analysing the raw data files: -44.0 ºC and –99.9 ºC. These data are easily removed using basic Excel macros (macros).
Then some periods of data are obviously moved a certain value in degrees either warmer or colder (actually almost always warmer). We call this phenomenon *temperature shift*. In the best case, the boundary between unshifted and shifted data is very clear. A simple macro makes it then possible to move a given range of data a chosen value in degrees (negative or positive). But generally this boundary is actually a period of time where, for example, one out of two measures is still correct. A macro was written, that makes it possible to define a period of time where some data are likely to be moved warmer. A boundary value is defined as well as a shift temperature equal to the presumed difference between the moved data values and the correct

values. All data superior to the maximum value are then moved colder by the amount of degrees of the shift temperature. The results from this data sorting, if not perfect, makes it possible to use data that else would be unusable (Figure 5-3)

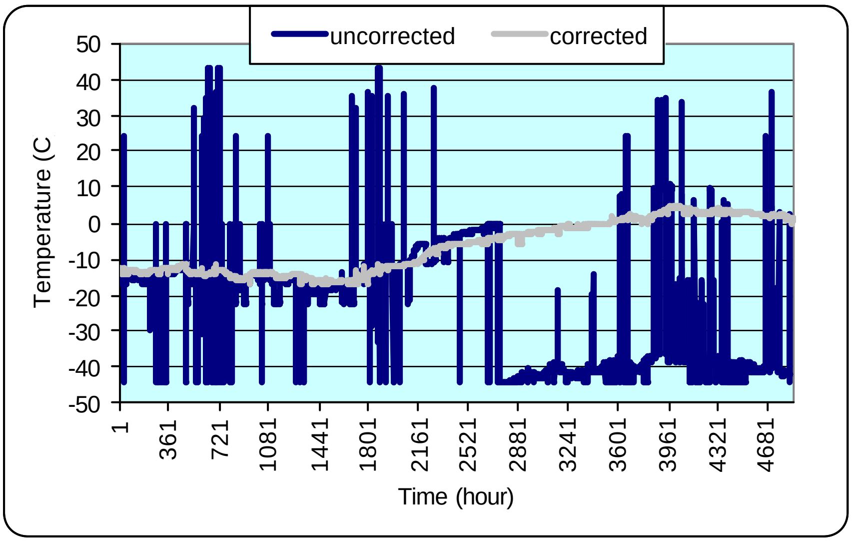



Figure 5-3 Example of data with *temperature shift* before and after sorting

Finally some single data are randomly wrong and are almost impossible to correct. A simple sorting macro was however programmed. This macro takes a maximum hourly variation as input. When a value is shown to differ from its preceding value by more than the defined maximum variation, it is set to the same value as this preceding value. This macro is especially useful at some depth under the surface, where high frequency variations of the air temperature do not penetrate and the maximum hourly variation generally ranges from 0.2 to 0.5 °C. It is important to systematically compare the results after running this macro with the original data. The error occurring most frequently is that the curve becomes constant over a long period. This is typically the case when data fluctuate alternatively over and under a mean value. The difference from one hour to the other is then the double of the difference between one value and the average value.

When it comes to air temperature series, the data from the NGI weather station are of good quality for the first years but degrade after a while. The reference used for estimating the quality of the air temperature data is the temperature series measured by DNMI at Sveagruva. A comparison of NGI and DNMI air temperature data for the year 1980 shows that DNMI data gives a good approximation of the air temperature at the permafrost measurements site (Figure 5-4).

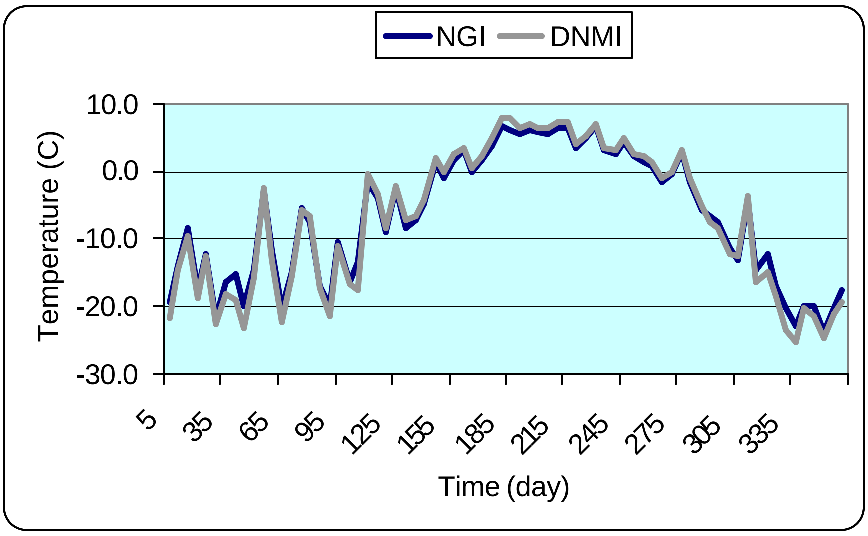



Figure 5-4 Comparison NGI/DNMI air temperature year 1980

### 5.3 Results

Processing of the data from the NGI weather station at Sveagruva shows that temperature series of the upper ground layer (down to 100 cm) are of good quality while under 100 cm, except from 800 cm, temperature series are of poor quality. At 150 cm and 200 cm depth, usable data is available for only 9 of 22 years. At 250 cm and 300 cm depth, the thermistors were obviously not working from 1982. Finally the data at 800 cm are missing for some years but can easily be interpolated as the temperature at this depth is slowly increasing, more or less linearly (Figure 5-5).

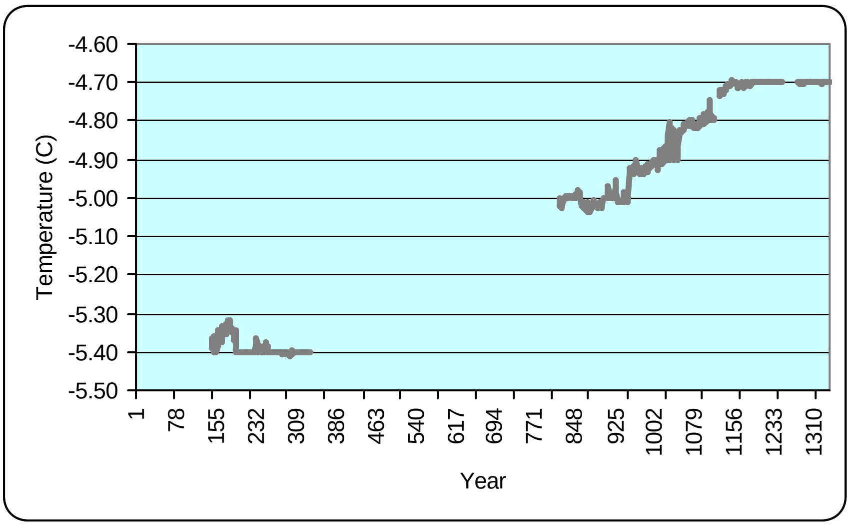



Figure 5-5 Temperature series at 800 cm depth

Snow depth were obtained by DNMI measurements for the whole period. These data show a significant variation from year to year in depth and duration of the snow cover in Sveagruva (Figure 5-6).

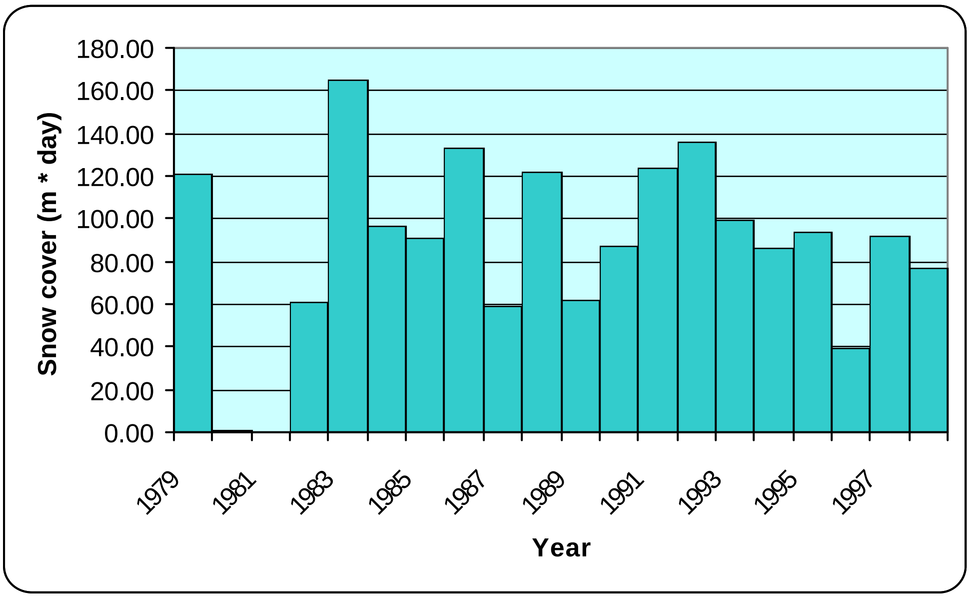



Figure 5-6 Snow cover in Sveagruva

Data on solar radiation were not sorted, neither were temperature data measured under the concrete surface. The reason for this is that the model used does unfortunately not allow to define a fluctuating heat flux as a boundary condition. Therefore it is impossible to simulate the effect of solar radiation throughout the year.

# Chapter 6 – Model

### *6.1 Software presentation*

The software used for modelling permafrost temperature is Geo-Slope International's TEMP/W. TEMP/W is a commercial finite element software product that can be used to model the thermal changes in the ground due to environmental changes or due to the construction of facilities such as buildings or pipelines. The software manages the whole modelling process from problem definition to problem solving and to results contouring and graphing.

TEMP/W is formulated to analyze ground temperature changes, including the effects of latent heat of fusion of ice. Hence, it has application in the analysis and design of facilities that are subjected to freezing and thawing temperature changes.

Finally, it is a 32-bit, graphical software product that operates under Microsoft Windows 95/98 and Windows NT/2000.

### *6.2 Parameters and boundary conditions*

The time is measured in days. The consistent set of units to use is therefore as follows :

- distances in metres (m)
- heat in kilojoules (kJ)
- temperature in Kelvin (K)

The material properties used are those presented in chapter 4 (note that the values given in chapter 4 are in S.I. units).

The variations of the snow cover cannot be modelled with the software. A constant snow cover is therefore used as an approximation. Basically the snow acts as an insulating layer from the air temperature in the winter, and it melts in the summer. The depth of this top surface layer is set to 25 cm for the testing year (1981) and thermal properties of the snow are taken to be that of drifted and compacted snow :

volumetric heat capacity : 500 $kJ \cdot m^{-3} \cdot K^{-1}$

thermal conductivity : 30 $kJ \cdot day^{-1} \cdot m^{-1} \cdot K^{-1}$

In the summer, the snow complete melting is simulated with a null volumetric heat capacity and a very high thermal conductivity (1000 $kJ \cdot day^{-1} \cdot m^{-1} \cdot K^{-1}$). The 25 cm deep top surface layer acts thus as a transparent layer for the air temperature, so that the air temperature is actually directly applied to the top of the ground (depth 0 cm).

However, for the snow simulation to work fully, we need to take the unfrozen water content function constant and null. This is done to avoid the simulated presence of water in the 25 cm deep top surface layer, which would modify the air temperature before it reaches the top of the ground.

Finally, the limit between winter and summer used for determining when the snow melts, showed after some testing to correspond to air temperatures superior to –8 ºC. For the testing year (1981), this is situated around the middle of May, i. e. the time when the sun begins to raise high in the sky and the snow rapidly melts in the valleys and along the shore. The limit between summer and winter corresponds to the middle of November (in 1981). On Svalbard the snow generally begins to come as soon as in september, however, due to strong wind, it takes some time before there really is a snow cover on the ground. Simulating the appearance of this cover in the middle of November seems therefore to be a good approximation of reality.

As described in the precedent paragraph, the boundary conditions are mixt (Dirichlet-Neumann). On top of the snow cover (z = 25 cm) and bottom of the permafrost (z = -800 cm), a temperature is applied while on the sides of the mesh, a null temperature gradient is applied.

The temperature on top of the snow cover is the measured series from DNMI. The DNMI station records air temperature three times a day (0700, 1300 and 1900). The model was run both with daily and 5-days average values, and the difference in the results was negligible (less than 1%). All simulations were therefore run with 5-days average values.

At the bottom of the permafrost, the temperature is almost constant (less than 0.1 ºC variation) during one year. The most convenient boundary condition is therefore a constant temperature. Moreover the heat flux at z= -800 cm is not constant throughout the year, so it is not possible to use any other type of boundary condition at the bottom of the permafrost anyway. In fact heat flux boundary conditions in TEMP/W can only be constant values, unlike temperature boundary conditions.

As we mentionned in chapter 5, the temperature at z = -800 cm raises slowly from 1978 to 1996, from –5.4 ºC to –4.7 ºC. This makes it difficult to run the model with a constant temperature as boundary condition at this depth. However the influence on the thermal regime of the ground of a 0.5 ºC difference of temperature on the bottom boundary condition was found to be negligible (less than 1%), at least over z= -300 cm (comparison between z= -800 cm and z= -300 cm is impossible since there are no thermistors betweeen these two depths).

### *6.3 Calibration*

The model was calibrated testing the influence of all parameters : snow depth, heat capacity, heat conductivity, heat flux, initial temperature, size of the mesh elements, heterogeneous model, volumetric water content, unfrozen water content.

The detailed results of this analysis are not detailed here but the main conclusion is that the ground temperature response mainly and greatly depends on its unfrozen water content function, making it necessar to have precise data for this function.

Furthermore, calibration is generally done with the very handy n-factor. A n-factor-like feature is proposed by TEMP/W, called “modifier function”. This feature was tested in order to increase the velocity of temperature raise in the ground during the spring. The effect was almost no change in this velocity but a raise in the maximum temperature reached at all depths. The modifier function is therefore not used in the final model.

### *6.4 Convergence*

The TEMP/W solver module gives information on the convergence of the calculations at each step. The convergence parameters used for the permafrost model are:

Maximum number of iterations : 50

Tolerance : 0.01%

Gauss regions / iteration : 4

Tests were made on steps where convergence was not reached. In these cases, convergence was never reached, neither by modifying the maximum number of iterations nor by modifying the tolerance value. The best way to correct convergence problems is therefore probably to sort air temperature series in order to avoid too important fluctuations.

Generally non-convergence is observed during the automn when the active layer begins to freeze. When running a simulation, one should systematically check that all steps converge, and if not the case, consider removing high frequency variations of the air temperature in the period of non-convergence. The properties of the snow could possibly also be changed in order to dispatch their variations on a wider range of temperatures.

# Chapter 7 – Results

## *7.1 One year modelling*

The permafrost model created with TEMP/W was first run on a one-year basis in order to calibrate the different parameters. The results from the one-year simulation are of low precision for the active layer during the summer while down in the permafrost, they are more coherent with measured data, for the whole year.
The low precision of the results in the active layer is illustrated in Figure 7-1, which compares measured data with simulated data for year 1981 at z = -20 cm.
While in the winter the simulated temperature series, Tsim, is in good accordance with measured temperature series, Tmeas ($|Tsim - Tmeas| \leq 1$ °C), during the thawing period, it increases too slowly and the difference between both series rapidly attains 5 ºC. At the end of the summer, this difference has reduced to 2 ºC. Finally, in December, the two series are again in good accordance.
The differences between measured and simulated data in the active layer is most probably due to the lack of precision of the unfrozen water content. Moreover the unfrozen water content function we use (Berggren, 1983) has been obtained from core samples taken under z= -100 cm. The unfrozen water content of active layer Svea clay is likely to be different from that of permafrost Svea clay and it would therefore be particularly interesting to take ground samples of the active layer for laboratory analysis.

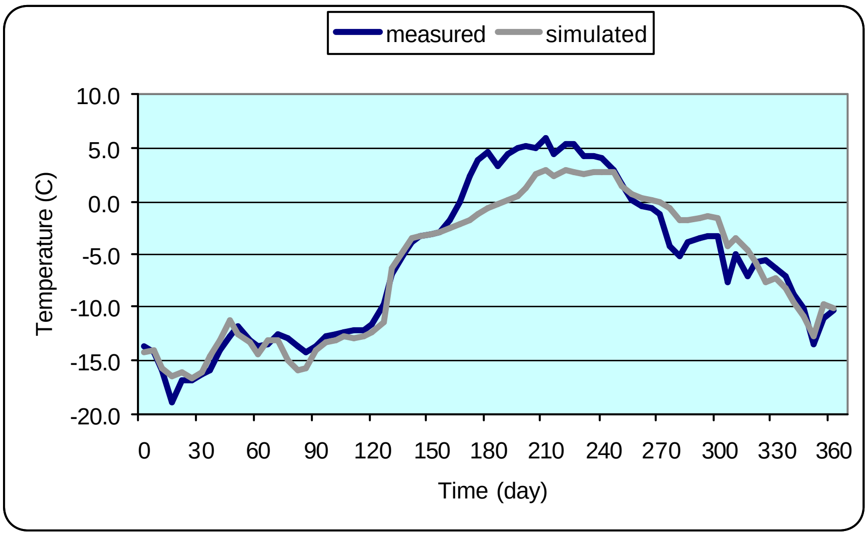



Figure 7-1 Temperature at z = -20 cm

In the permafrost, i. e. under z ≈ -80 cm, however, the precision of Tsim is acceptable (Figure 7-2). During the whole year, the difference between Tsim and Tmeas at z = -200 cm never exceeds 0.5 ºC. At other depths, the difference is slightly higher. The maximum registered differences are 1.6 ºC at z = -100 cm, 1.2 ºC at z = -150 cm and 0.9 ºC at z = -250 cm (data for z = -300 cm not available).

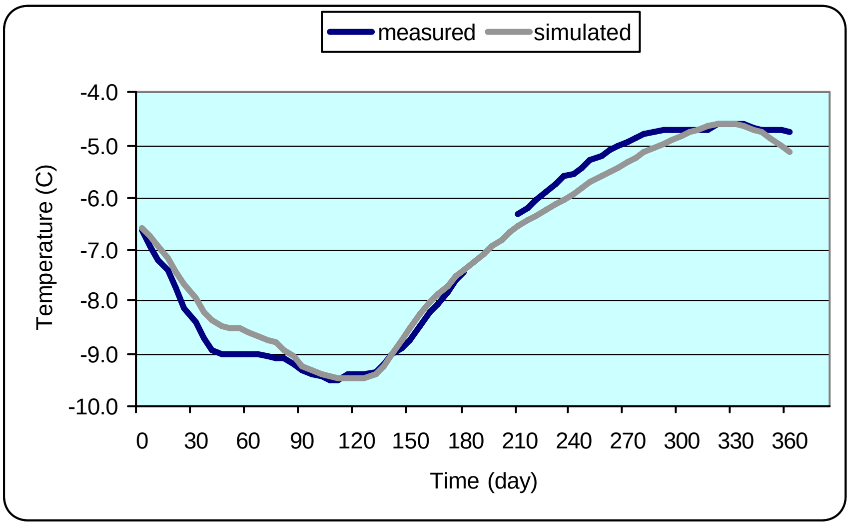



Figure 7-2 Temperature at z = -200 cm

Since the precision of the results from the simulation is low in the active layer, a model of permafrost alone has been tested. Basically this model is the same as the global one, only the upper 100 cm of ground (and the snow cover) have been removed and the measured temperature at z= -100 cm is used as upper boundary condition. The results from this simulation are closer to measured values, as we can see in Figure 7-3. The good precision (0.2 ºC) obtained with this simulation confirms the correctness of the permafrost material properties measured by Berggren (1983).

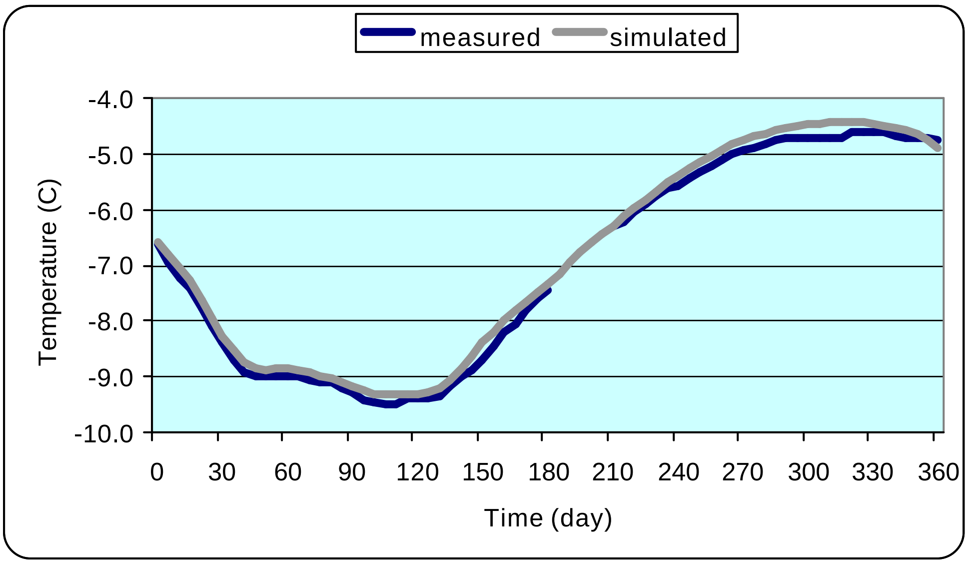



Figure 7-3 – simulation without active layer

### *7.2 Long-term modelling*

Since air temperature series of good quality are available from DNMI, it is possible to run the simulation on the whole period of measures, i. e. from 1979 to 1998. When working with long temperature series in TEMP/W, it is unthinkable to type all the data under TEMP/W. The procedure to use is to open the *.tem file in a simple text editor (e. g. Notepad) and to paste the data series from e. g. an Excel file. Before proceeding to this cut-and-paste operation, the format of TEMP/W files should be thoroughly checked in the help topics of TEMP/W.

The results at z = -100 cm from the simulation on a 20 years period (1979 to 1998) are presented in Figure 7-4. Extreme values differ significantly from Tsim to Tmeas. In the summer, we have up to 1 ºC difference and in the winter, up to over 3 ºC. The summer difference is in accordance to what we had for one year simulations. For the differences between measured and simulated minimum temperature in the winter, a further analysis shows that it is directly connected to the thickness of the snow cover. Indeed if we consider the second year, we have a difference of over 3 degrees between the two minima. If we take a look at the chart of the snow layer for that winter (1979 to 1980), we see that the amount of m · days of snow is extraordinarily low (1.16) compared to the average value (87.3). With almost no snow cover, the cooling of the ground is much higher than for a usual winter.
This observation highlights the first approximation made in the 20 years model. Indeed a constant winter snow cover is used while in reality the thickness of the snow cover varies during one winter as well as from one year to another (Figure 5-6).
The second important approximation made in the 20 years model is the use of a constant temperature of –5 ºC at the bottom of the element mesh (z = -800 cm). In reality we have discussed the fact that the temperature at this depth follows a slow linear increase. The ideal would of course be to build a model on which a 20 years simulation predicts the same slow linear warming trend as observed in reality.
The warming trend, on the 20 years of existence of the Sveagruva weather station, is in accordance with the increase of the thawing index and the decrease of the freezing index during this period (Instanes, personal communication). Whether this increase is local in time or confirms expectations of global warming is not the aim of this thesis to discuss. It can however be observed that equivalent trends (both warming and cooling) were observed earlier in the last century, on Svalbard (Instanes, personal communication). The installation by A. Instanes of a new, deeper (z = -1000 cm) thermistor in Sveagruva in September 2000 is a first step in determining whether the trend observed results from a global warming or not.

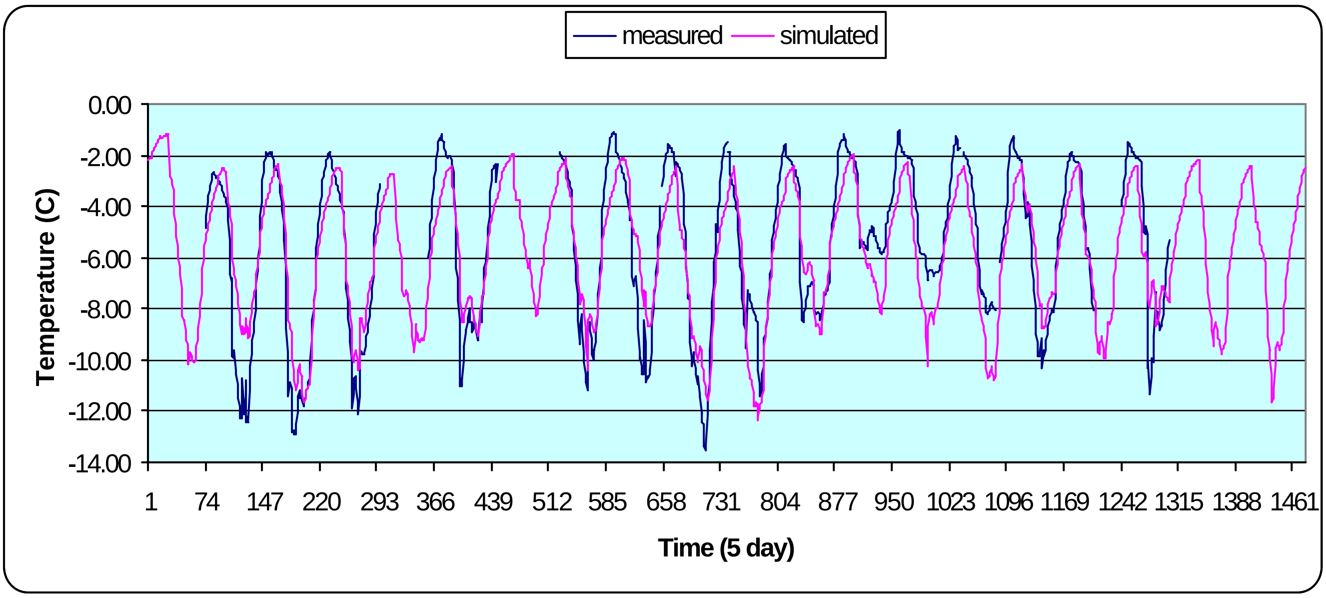


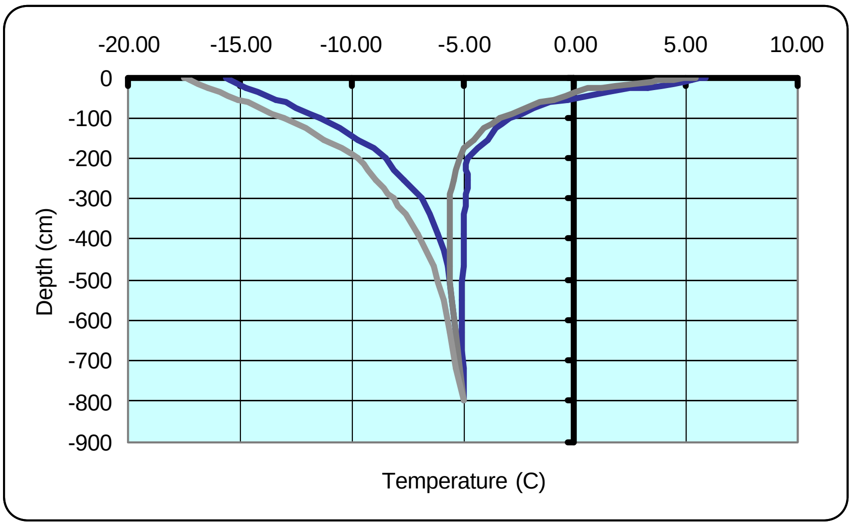

Figure 7-4 Temperature at z = -100 cm

### *7.3 Climate change scenarios*

As discussed in the precedent paragraph, the model built for simulating the ground thermal regime in Sveagruva is not so well adapted for long-term modelling as it is for modelling on a yearly basis. Climate change scenarios were, however, applied to the model in order to get an idea of the consequence of a change in the mean annual air temperature. Two situations were considered : 2 degrees warming and cooling of the mean annual air temperature on a period of 50 years. For each situation, both a step change and a linear change were simulated.

After 50 years, the difference between each scenario (step or linear change) is negligible. The temperature change is thus small enough to let the ground accomodate to it each year. As a result we can restrict our analysis to only one scenario.

Figures 7-5 and 7-6 below show the change in the whiplash curve resulting from a climate change. The light gray curves are the results from the simulations while the dark ones are measured values from 1978.



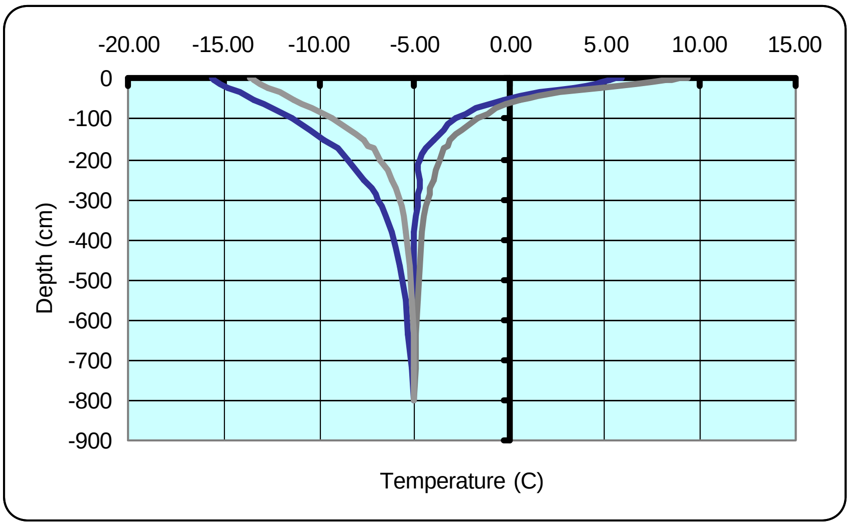




Figure 7-5 Whiplash cooling 50 years

Figure 7-6 Whiplash warming 50 years

The change in air temperature modifies the thermal regime in the ground and, thus, the depth of the permafrost table. Unfortunately the model used gives, on a one year basis, a depth of the permafrost table of 45 cm while in reality, it is about 80 cm. However if we consider that the change in the depth of the permafrost table is less than the double of that predicted by the model, we can assess, in first approximation, a change of  25 cm of the depth of the permafrost table.

# Chapter 8 – Conclusion

The model built for simulating permafrost response to annual variations of meteorological data is not considered precise enough to draw a conclusion as to the effect of a change of the mean annual air temperature (MAAT) in the Arctic. Only a first estimation can be made. This estimation, based on the simulation of a raise of 2 ºC of the MAAT during a period of 50 years, shows that the thickness of the active layer would increase by 25 cm from 80 cm today (permafrost degradation). Such an increase would not threaten constructions on Svalbard that, as a rule, are built on 12 m deep piles.

The imprecision of the model is due to the lack of data on the behaviour of the active layer. Additionaly the software used does not allow precise enough simulation, as neither variable snow cover nor variable solar radiation can be included in the model.

The behaviour of the active layer is difficult to model. It is therefore important to have several thermistors in this layer, while in the permafrost thermistors may be more scattered, as is the case at the NGI station of Sveagruva. Moreover thermistor failure is often observed and the data measured should be analysed on a regular basis in order to track such problems and correct them rapidly.

The thermistors at the NGI station of Sveagruva are not placed deep enough to study past climate changes, however the data measured at the bottom of the borehole, e. g. 8 m, can probably be used to study present climate change. The installation of a new thermistor at 10 m depth these days will increase the precision of such a study.

The analysis of the ground samples that will be taken during the same field session will hopefully bring valuable data for the further calibration of the model built during this thesis.

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