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Surface and subsurface conditions in permafrost areas – a literature review

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Preface

This study, done during the autumn of 2001, is based on available literature and the author has tried his best to keep the facts fully objective and purely technical or scientifical. However, in areas where limited literature is available (e.g. concerning subjects such as taliks, volatile gas hydrate venting, etc) it is possible that the quoted authors', my reviewing colleagues', or my personal views are stated in a way of leading a scientific discussion.

Acknowledgements

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Professors J Åkerman at Lund University, M Seppälä at Helsinki University, and Sylvi Haldorsen at the Agricultural University of Norway together with Mr T Eliasson at SGU, Gothenburg are acknowledged for kindly sharing some of their available literature, which certainly made some papers and information available for me that otherwise would have been hard to assess.

Abstract

This report contains a summary of some of the information within existing technical and scientific literature on permafrost. Permafrost is viewed as one of the future climate driven process domains that may exist in Scandinavia, and that may give rise to significantly different surface and subsurface conditions than the present. Except for changes in the biosphere, permafrost may impact hydraulic, mechanical, and chemical subsurface processes and conditions. Permafrost and its influences on the subsurface conditions are thus of interest for the performance and safety assessments of deep geological waste repositories.

The definition of permafrost is "ground that stays at or below 0°C for at least two consecutive years". Permafrost will effect the geological subsurface to some depth. How deep the permafrost may grow is a function of the heat balance, thermal conditions at the surface and within the ground, and the geothermal heat flux from the Earth's inner parts.

The main chapters of the report summaries the knowledge on permafrost evolution, occurrence and distribution, and extracts information concerning hydrology and mechanical and chemical impacts due to permafrost related conditions.

The results of a literature review are always dependent on the available literature. Concerning permafrost there is some literature available from investigations in the field of long-term repositories and some from mining industries. However, reports of these investigations are few and the bulk of permafrost literature comes from the science departments concerned with surficial processes (e.g. geomorphology, hydrology, agriculture, etc) and from engineering concerns, such as foundation of constructions and pipeline design. This focus within the permafrost research inevitably yields a biased but also an abundant amount of information on localised surficial processes and a limited amount on regional and deep permafrost characteristics.

Possible conclusions are that there is little known on permafrost in hard bedrocks at the depths of relevance for spent nuclear fuel repositories as well as of permafrost characteristics of importance for the performance and safety assessments of such repositories. Some important observations are though that large lakes and rivers are plausible locations for taliks and may under certain conditions (e.g. glacial fore-fields) be subjected to large flows from deep bedrock. Salt rejection and hydrate concentrations are both likely chemical effects of permafrost. These effects are found at deep locations in contemporary permafrost regions, mechanical effects of permafrost are however surficial.

Sammanfattning

Den här rapporten summerar en del av informationen i befintlig teknisk och naturvetenskaplig litteratur om permafrost. Permafrost bedöms vara ett av de framtida klimatdrivna process-tillstånd som kan komma att existera i Skandinavien, och som kan innebära väsentligt ändrade förhållanden både på och under markytan jämfört med dagens. Förutom att påverka biosfären kan permafrost komma att påverka hydrauliska, mekaniska och kemiska processer och tillstånd. Av denna anledning är permafrost av betydelse inom funktions- och säkerhets-analyser för ett djupförvar.

Permafrost är definierat som ”mark som är vid eller under 0°C under åtminstone två efterföljande år”. Hur djupt permafrosten kan växa är en funktion av värmebalansen, termiska egenskaper och tillstånd i marken samt den geotermiska gradienten.

Rapportens huvudkapitel summerar kunskapsläget inom permafrosts utveckling, uppträdande och utbredning både i yta och mot djupet. Vidare behandlas information om hydrologi samt mekaniska och kemiska effekter på grund av permafrost.

Resultatet av en litteraturstudie beror alltid av tillgänglig litteratur. Vad det gäller permafrost finns en del litteratur från forskning kring djupförvar och från den arktiska gruvindustrin. Emellertid är denna information begränsad och den stora informationsmängden kommer från naturvetenskapliga institutioner som studerar ytnära processer (geomorfologi, hydrologi, agrikultur, m fl) och från ingenjörsområden såsom grundläggning och design av olje- och gasledningar. Denna fokus inom permafrostforskningen leder oundvikligen till en snedfördelad och riklig information inom lokala ytnära processer och tillstånd och en betydligt mindre kännedom om regional och djup permafrost.

Möjliga slutsatser är att endast lite är känt om permafrost i kristallin berggrund på de djup som är aktuella för djupförvar för använt kärnbränsle och de egenskaper hos permafrost som är av betydelse för funktions- och säkerhetsanalys av sådana förvar. Några viktiga iakttagelser är dock att stora sjöar och vattendrag är potentiella platser för taliks och att dessa under vissa förhållanden (t ex framför en glaciär) kan leda stora flöden från djupa delar av berggrunden. Utfrysning av salt och en koncentrering av hydrater är troliga kemiska effekter av permafrost. Båda dessa effekter återfinns på stort djup i dagens permafrostområden, mekaniska effekter är emellertid ytnära.

Content

1 Introduction

Surface and subsurface conditions in the geosphere depend on the governing climate and its local variations. Further, the climate has been shown to have varied through time (e.g. /Holmgren and Karlén, 1998/; and described in e.g. /Andersson, 1998; Morén and Påsse, 2001/). The physics behind these global climate changes predict that also the future climate will vary /e.g. Boulton and Payne, 1992/. The time scale of the lifecycle of a deep repository is in the order of 10,000 to a million years. In this time scale the climate in the Swedish region of the Earth will probably experience both glacial and boreal conditions, possibly more than one time. In between these conditions there will be both permafrost and high sea-levels effecting the conditions that have an impact on the deep repository.

Therefore permafrost is viewed as one of the key surface boundary domains /SKB, 1999; Boulton et al, 2001/ that may exist in large parts of Scandinavia, and that may give rise to significantly different surface and subsurface conditions than that of the present. Except for changes in the biosphere, permafrost may impact hydraulic, mechanical, and chemical subsurface processes and conditions. Permafrost and its influences on the subsurface conditions thus are of interest for the performance and safety assessments of a deep geological waste repository.

Up to the present climate research has not been able to give very detailed information about the preglacial (i.e. before the last glaciation, the Weichselian) conditions in Sweden. This is mainly due to the amended surface conditions that resulted after the erosional processes following the coverage by the Weichselian ice sheet. However, more is known on the postglacial (Holocene) bedrock environment.

1.1 Definitions

Permafrost is frozen ground that stays frozen for at least two consecutive years /Williams and Smith, 1989/. Herein frozen is defined as being below 0°C, that is, within permafrost ground there may exist liquid water due to e.g. chemically altered freezing point, pressure induced changes etc. Since, the use of frozen more often refers to the presence of ice; the notion *cryotic ground* is used for ground that is below 0°C and the permafrost definition is: ground that stays at or below 0°C for at least two years /French, 1996; IPA, 1998/. The presence of permafrost is likely to affect the geological subsurface to some depth. How deep the permafrost may grow is a function of the heat balance, thermal conditions at the surface and within the ground, and the geothermal heat flux from the Earth's inner parts.

Permafrost is a part of the periglacial environment. Modern usage of the *periglacial* concept refers to a broad range of non-glacial cold conditions regardless of their proximity to a glacier /Washburn, 1979/. For most purposes, it is sufficient to define periglacial environments as those where frost action or permafrost processes dominate. Today, periglacial environments occupy more than 20 per cent of the continental land surface; however, this is not a static percentage. During the last glacial maximum an additional of approximate 25 per cent /French, 1996/ may have experienced periglacial conditions. In addition to periglacial permafrost, permafrost can also occur outside the defined periglacial regions, for example beneath glaciers. The ground in front of a glacier is, however, well defined within the periglacial concept but these glacial fore field grounds may also be described as proglacial environments.

On top of a permafrost layer, an *active layer* exists (see Figure 1-1). The thickness of the active layer may be of centimetre to metre scale. In this layer repeatedly freezing and thawing occur during a year cycle. For some time of a year parts of an active layer will be over-saturated with liquid water and for some other time the ground will be frozen and partly impervious for groundwater formation and flow.

The active layer is by the most common definition defined as a layer in which the liquid water freezes to ice and then thaws again /van Everdingen, 1976/. However, the contrasting definition, solely on temperature (following the same definition as on permafrost), is used in parallel /Burn, 1998/.

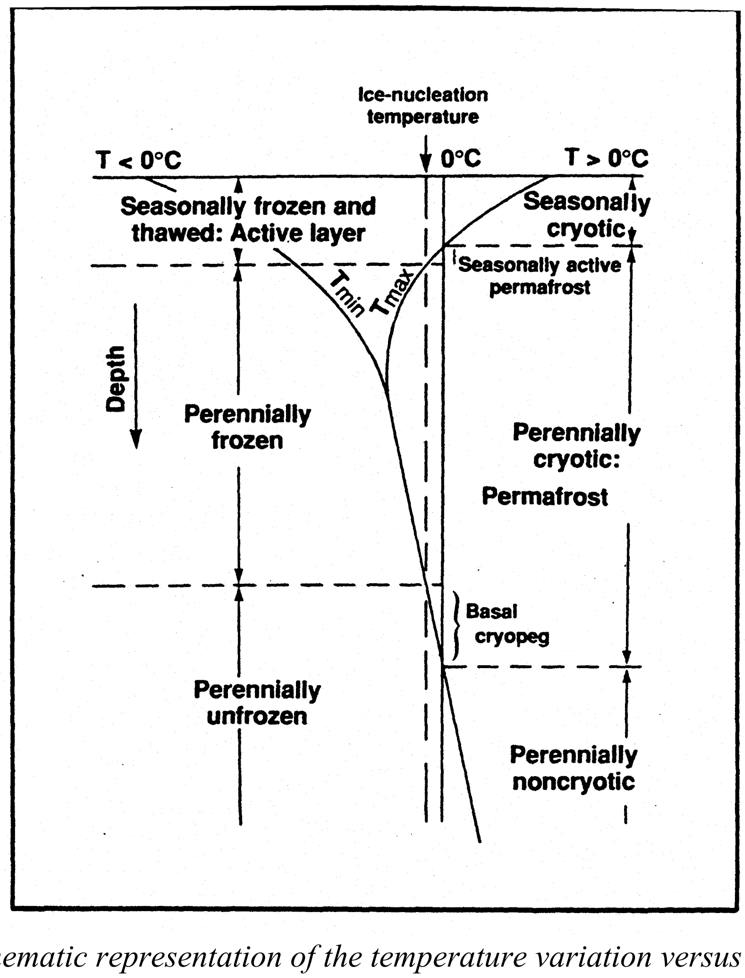

Figure 1-1. *Schematic representation of the temperature variation versus depth with illustration of the active layer as defined by /van Everdingen, 1976/, including the temperature related definitions of cryotic and permafrost /from van Everdingen, 1990/.*

In addition to the active layer, which stays unfrozen for parts of the year grounds that stay perennially unfrozen exist within the permafrost environment. These unfrozen grounds may exist due to an anomaly caused by e.g. thermal, chemical, hydrological, or geological conditions. These unfrozen grounds are defined as *taliks* /IPA, 1998/. However, herein the notation of talik is used for a water body that is not a part of the active layer or the non-saline waters existing beneath the permafrost, or not being a cryopeg. This use of the notation of talik is in agreement with the literature reviewed.

Ground waters that stay unfrozen due to altered or initially saline composition and which are perennially cryotic (temperature below 0°C) are called *cryopegs*. These water bodies could be basal cryopegs that exist between the bottom layer of the frozen permafrost and the waters existing in the sub-permafrost system. However cryopegs could also be marine, or found within the frozen ground as isolated unfrozen water bodies. Cryopegs are a part of the permafrost.

Permafrost regions are in general classified into being either *continuous* or *discontinuous* (cf Figure 1-2). In continuous areas frozen ground is present everywhere except for localised taliks /French, 1996/ existing beneath lakes and rivers. In discontinuous regions areas of unfrozen ground separate the permafrost /French, 1996/. On the warmer fringe of the discontinuous permafrost zone isolated 'isles' of frozen ground is present and this part of the permafrost is often defined as *scattered* /Williams and Smith, 1989/ or, herein, as *sporadic* permafrost, following a frequent use in scientific papers.

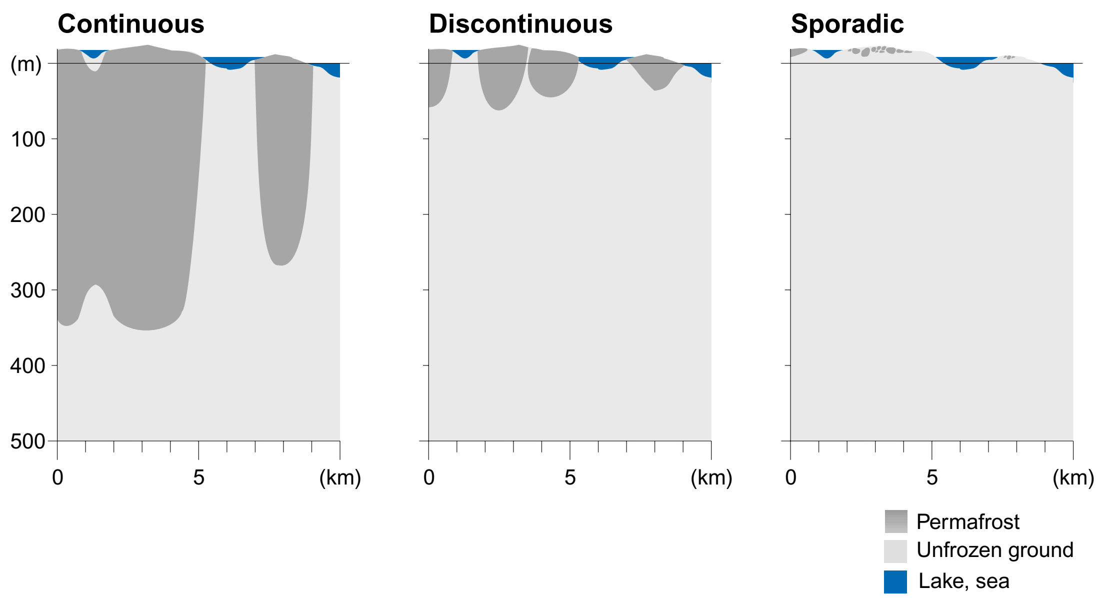

Figure 1-2. *Illustration describing the definitions of continuous, discontinuous and sporadic permafrost /modified after French, 1996/.*

1.2 Scope of the study

This literature review aims at a compilation of the state-of-the-art knowledge in the permafrost research. However, the review is restricted in the sense that the main focus is on permafrost which may effect the safety and performance of a deep repository within the bedrock. Further, this review is a compliment to previous reviews by SKI /McEwen and de Marsily, 1991/ and SKB and POSIVA /Gascoyne, 2000; Ahonen, 2001/.

The main objective of the report is to give guidance concerning the remaining tasks related to permafrost identified in SKB:s safety assessment SR97 /SKB, 1999; SKB, 2001/:

- Occurrence and depth distribution
- Hydraulic conditions
- Chemical conditions
- Mechanical implications
- Biosphere conditions.

The results of a literature review are always dependent on the available literature. Concerning permafrost there is some literature available from investigations in the field of long-term repositories and some from mining industries. However, reports of these investigations are few and the bulk of permafrost literature comes from the science departments concerned with surficial processes (e.g. geomorphology, hydrology, agriculture, etc) and from engineering concerns, such as foundation of constructions and pipeline design. Further, in the wake of the enhanced question of anthropogenic greenhouse gases, new research concerned with permafrost degradation due to climate warming is also frequent in the available literature.

This focus within the permafrost research inevitably yields a biased but also an abundant amount of information on localised surficial processes and a limited amount on regional and deep permafrost characteristics. A relevant but limited compilation on permafrost depth information is presented. However, the presented information concerning mechanical implications has all been related to surficial processes and soil mining and may therefore have a weak importance for the knowledge of the deeper bedrock conditions. The biosphere conditions are just briefly touched upon if the biosphere impacts on the thermal or hydraulic processes described.

2 Permafrost evolution

The presence of permafrost in ground depends on the ground temperature, which is controlled by a heat exchange across the atmosphere/ground boundary layers /Williams and Smith, 1989; French, 1996; Smith and Riseborough, 1996/. The heat exchange is governed by air temperature, which in turn depends on both regional and local climate conditions. However the heat exchange also depends on the heat flux from the Earth's interior. Properties and conditions along the atmosphere/ground boundary, such as thickness of layers, air temperature, air moisture, snow cover, vegetation cover, and ground characteristics, etc (cf section 2.3 and section 2.4) affect the heat flux and thus control the resulting ground temperature. Because of microclimate features such as those described above the ground surface temperature usually differs by several degrees from corresponding air temperatures /Brown, 1974/. A value of approximately 3.5°C warmer ground surface temperature could be used as an average figure, but the normal scatter falls within a range of 1 and 5.5°C.

2.1 Temperature definitions

The air temperature varies throughout the year. As a consequence one is normally forced to work with the concept of the mean annual air temperature (MAAT). This mean temperature is a climatic parameter normally defined at a height of 1.5 to 2 metres above ground surface /Liljequist, 1970/.

The ground temperature profile versus depth has a transient nature and especially the surficial regions have significant temperature variations throughout the year (cf Figure 1-1). Therefore a mean annual ground temperature (MAGT) is often used to define the ground temperature. The MAGT can be defined for any depth; if the ground surface temperature is specified this is stated as mean annual ground surface temperature (MAGST). However, the exact location of the ground surface interface is difficult to assess. Therefore the use of MAGT is preferable.

At a certain depth in the subsurface, there will be no apparent effect of the temperature variation at the surface and from this depth and down there will be equilibrium between the internal heat flow and the average conditions in this surficial layer /e.g. Gold and Lachenbruch, 1973/. This depth of no annual temperature variations is therefore used to describe the thermal equilibrium within the ground.

2.2 Heat exchange balance

The processes controlling the heat exchange at the atmosphere/ground boundary are the *net exchange of radiation* (Q^*) between the surface and the atmosphere, the *transfer of sensible heat* (Q_H), due to turbulent motion of air, and of *latent heat* (Q_{LE}, evaporative/condensative heat flux); the last part in the heat exchange balance is the *conduction of heat* (Q_G) into the subsurface.

$$Q^* = Q_H + Q_{LE} + Q_G$$

where

Q^* = Net radiation

Q_H = Sensible heat

Q_{LE} = Latent heat

Q_G = Conductive heat

The net exchange of radiation depends on the fluxes of solar short-wave (*K*) and terrestrial long-wave (*L*) radiation. But also local conditions such as, topography and vegetation affect the net radiation /Williams and Smith, 1989/.

$$Q^* = K\downarrow(1-\alpha) + L\downarrow - L\uparrow$$

where

Q^* = Net radiation

$K\downarrow$ = Incoming short - wave radiation

α = Coefficient of surface albedo

$L\downarrow$ = Incoming long - wave radiation

$L\uparrow$ = Outgoing long - wave radiation.

The incoming short-wave radiation is primarily depending on latitude, time of year and day, and weather conditions such as cloudiness. The short-wave radiation that reaches the ground is reflected to a certain degree, dependent on the surface's albedo (α) value; the higher the albedo the colder surface temperature will result.

Table 2-1. Typical albedo (α) values for some surfaces /based on Williams and Smith, 1989/.

Ground surface	Typical albedo (α) values
Water	~ 0.10
Lichen heath (Peat)	~ 0.20
Stoney plain	~ 0.25
Snow	~ 0.80

The Earth surface further gains heat from atmospheric long-wave radiation, which is absorbed short-wave radiation emitted as long-wave radiation. This incoming long-wave radiation is a climatic factor, influenced by the atmospheric humidity and cloud conditions, but also of other constituents of the atmosphere /Paterson, 1994/. The ground surface also emits radiation, depending on the emissivity, a local thermal parameter, and the temperature of the surface.

$$L\uparrow = \varepsilon k T_s^4$$

where

$L\uparrow$ = Outgoing long - wave radiation

ε = Coefficient of emissivity

k = Thermal constant, dependant on local parameters

T_s = Surface temperature.

For most purposes the Earth surface can be viewed as a “black body” and the emissivity and the thermal constant (k) can be substituted by Stefan-Boltzmann’s constant.

Due to the Earth axial tilt, little solar energy is received in the Polar areas of the Earth, and in general a strong net loss of energy results in a surface cooling /Mai and Thomsen, 1993/. On average the radiation balance is regarded as negative on latitudes of 35° and higher /Liljequist, 1970/.

The sensible heat is described with:

$$Q_H = -\rho c_p K_H \, \partial T / \partial z$$

where

Q_H = Sensible heat flux

ρ = Density (of air)

c_p = Specific heat capacity (of air)

K_H = Turbulence transfer coefficient

$\partial T / \partial z$ = Atmospheric temperature gradient.

The turbulence transfer coefficient (K_H) depends on the local wind conditions and the surface roughness. A rougher Earth surface, such as a forest, will tend to be cooler /Williams and Smith, 1989/ compared to a smoother environment, such as an open field.

The latent heat process is complex and contains sub-processes such as heat exchange in relation to thawing of snow and ice, heat exchange during the transfer of water to vapour along the ground surface and vice versa, as well as heat exchange during freezing of water. The rate of evaporation (evapotranspiration) depends on climatic variables such as temperature, wind, and moisture content /Williams and Smith, 1989/. For the case of vaporisation the latent heat is:

$$Q_{LE} = -\rho L_v K_v \, \partial q / \partial z$$

where

Q_{LE} = Latent heat flux

ρ = Density

L_v = Coefficient of latent heat (of vaporisation)

K_v = Turbulent transfer coefficient (of vapour)

q = Specific humidity (in the air)

$\partial q / \partial z$ = Moisture gradient.

Cyclic warming and cooling of the ground have the tendency to cool the ground after each event of thawing /Lunardini, 1993/ and vice versa. Therefore the time of permafrost evolution will be greatly lengthier than the time predicted by pure conductive process. The latent heat effect of the freeze/thaw cycles has however no significant impact on the final total thickness of the permafrost.

Heat transfer within the ground can occur through conduction, convection, and radiation. The heat transfer due to radiation depends on the thermal gradient but also on the amount of stored heat. For most cold conditions the radiation part can be neglected /Sundberg, 1988/. Further in the cold environment, convection can only be of importance when groundwater fluxes are significantly large. As a consequence, in general, only conduction is viewed as important in permafrost evolution /Williams and Smith, 1989/. The conduction of heat in the ground depends on the thermal properties of the ground matter and on the temperature difference. The conduction of heat into the ground is often neglected in an annual perspective due to that the induced excess of spring and summer heat is balanced by the release of stored heat to the atmosphere during the autumn and winter seasons /Williams and Smith, 1989/. However, there is also a constant flux of heat from the Earth's interior (cf section 0).

Thermal problems are frequently investigated with the aid of numerical models, a review of such methods are presented by e.g. /Alexiades and Solomon, 1993/. /Romanovsky et al, 1997/ evaluated three different numerical models frequently used in permafrost modelling. They concluded that for these models to yield results that show a good correspondence with analytical as well as field data, their solution scheme needs to use small time- and depth steps.

/Malevsky-Malevich et al, 2001/ used a one-dimensional thermal model to investigate areal distribution of active layer thickness in Siberia and the response of the active layer to climate change. The resulting model thickness ranges of the active layer were found to yield good correspondence with field measured data. /Malevsky-Malevich et al, 2001/ further compared their simulated temperatures at depth with field measured data and again illustrated a good robustness of their code. However, they also concluded on a need for improvements, especially in the surface layer where interactions of atmosphere, vegetation and the ground are not well understood.

Numerical simulations of bedrock /Kukkonen and Šafanda, 2001/ indicate that the bedrock responds quickly to variations in ground temperature. These simulations demonstrated a migration of the permafrost front of approximately 0.1 metre per year when the surface temperature was changed by 1.5 degrees. /Delisle, 1998/ used a two-dimensional numerical model for simulation of permafrost growth and decay during the last 50,000 years at a hypothetical site in Northern Germany. Using linearised time-dependent temperature data, based on mean annual air temperatures given by /Vandenberghe and Pissart, 1993/, Delisle concluded that at no time during the modelled time cycle was the temperature sufficiently low and stable long enough to create a thermal equilibrium. The total permafrost thickness stayed less than 120 metres at all times.

As part of an ice sheet simulation for the Weichselian, /Boulton and Payne, 1992/ produced permafrost depth distributions along a transect from the Norwegian continental margin down to Northern Poland. This transect passed over the central Swedish highlands (the area of Jönköping). The model result showed that a permafrost situation developed prior to the ice sheet overrode the area. The result indicated that the depth of this permafrost was between 80 and 240 metres. Further the insulating ice sheet in the model of Boulton and Payne caused the permafrost to melt as soon as the ice sits on top of the region and within some thousands of years the permafrost has disappeared.

2.3 Climate

The climate varies on all time scales and in response to both random and periodic as well as possibly anthropogenic forcing factors /e.g. Goodess et al, 1992/. The climate can be defined as the normal condition within the atmosphere /Liljequist, 1970/. The governing processes in the atmosphere depend on a number of meteorological variables. Of these temperature, precipitation, wind, and their seasonal distribution /Washburn, 1979/ are the most important climatic parameters governing the evolution of permafrost and its associated processes.

Water input to permafrost areas has one major source, namely precipitation /e.g. Nelson et al, 1993/. Both forms of precipitation (rain and snow) and the frequencies and amplitudes with which it falls are of importance for the ground water content, groundwater recharge, etc. The ground water content effect the ground matters thermal properties, but more importantly does a snow cover impact significantly on the surface albedo.

Precipitation alone does not yield the available amount of water. The wind is responsible for redistribution of snow and soils. Air moisture migrates both through diffusion processes and advective processes. Therefore, in the air the wind is responsible for the depletion of moisture. Everywhere at the Earth surface there is ongoing evaporation /Liljequist, 1970/. The temperature, air moisture, and the wind govern evaporation from a free water surface. Also the sublimation depends on the dryness of air relative to the snow surface. In certain regions of North America /Woo, 1986/ the sublimation and evaporation of available melt waters are so intense that snow covers may be depleted without yielding any surface run-off or groundwater recharge.

Additionally, the effective heat loss from the ground is affected by the wind conditions. The turbulent air motion in the vicinity of the ground surface governs the sensible heat flow.

Within such a complex system as the climate there are always positive and negative feedbacks, with responses hard to quantify. At present the Earth experiences a climate warming, which would cause permafrost to melt and release carbon /Tarnocai, 1999/ and other greenhouse gases to the atmosphere and hence possibly enhance the climate warming /Anisimov and Nelson, 1996/ and its associated changes in precipitation and wind conditions. In a permafrost region, a climate warming of any kind will likely be accompanied with an associated change in the precipitation regime /Gavrilova, 1993; Nelson et al, 1993/. Many palaeoclimate simulations that have implied higher historical temperatures than what is congruent with palaeoclimatic proxies illustrate the importance of the feedback mechanisms. /Renssen et al, 2000/ performed modelling where they used a ground surface boundary more realistic for a permafrost environment, that is wet and frozen. With this forced use of a feedback mechanism their results came to show a much better fit, than most other simulations, with the climatic records that do exist.

2.4 Ground conditions and parameters

Features of the microclimate at a site usually are more important in permafrost development and occurrence than are macroclimate effects /Williams and Smith, 1989/. Even though local conditions (e.g. thermal, hydrologic etc) are linked to the atmospheric climate the significance of the air conditions are moderated by ground surface processes occurring within the boundary layers of snow, vegetation, organic matter in soils, and ground matter (regoliths/rocks) itself.

2.4.1 Snow cover

The distribution of a snow cover and the differences in snow depth are of great importance in the permafrost environment /Brown, 1974; Åkerman, 1980; King and Seppälä, 1987; French, 1996/. In general, a strong positive correlation between atmospheric and ground surface temperature exists /Isard and Schaetzl, 1995/. In snow free periods, such as July and August, Isard and Schaetzl observed a correlation coefficient of 0.99, however during the winter period, with a snow cover, their observed correlation decreased significantly and during February the coefficient was down to 0.28. It should be noted that these correlation coefficients are based on one investigation area only.

Viewed from the perspective of permafrost evolution the moderating effect on the ground temperature due to a snow cover is the most important result of snow /Williams and Smith, 1989/. According to /Seppälä, 1997/ the temperature at the snow/ground boundary under a one-metre thick snow cover is constant and approximately 0°C. Methods for estimation of the insulating effect of a snow cover on ground temperature have been established /e.g. Nicholson and Granberg, 1973; Linheng and Youwu, 1993/ however, no exact solution is available. /Desrochers and Granberg, 1988/ illustrated the complexity of heat transfer in a snow cover and concluded that a great amount of factors complicate the relationship, which is however mainly based on snow cover thickness. Additional factors of importance for the thermal conductivity of a snow cover are the geographical location of the snow cover, the density of the snow pack and its different layers /Linheng and Youwu, 1993/.

/Bakkehøi and Bandis, 1988/ reported on a strong relationship between time, depth of the snow cover, and the depth of the active layer. They observed that where the snow cover performed insulation to the extreme cold temperatures in winter the thawing depth during summer was significantly deeper than in areas that had stayed snow free throughout the entire winter period.

2.4.2 Topography

Topographical highs and lows have a significant impact on the air temperatures /e.g. King and Seppälä, 1987/. The air temperature could reach significantly lower values at the bottoms of topographical sinks. The lower temperatures in topographical sinks do not by necessity imply permafrost growth, this since these topographical sinks also function as snow traps, compared to the wind blown summit areas, and in such a case will a snow cover effectively insulate the ground matter.

The slope direction and azimuth impact on the net exchange of short-wave radiation. As a result north slopes receive much less radiation than do south slopes. Therefore thickness and in general occurrence of permafrost is larger on the north side of a summit than on the south. This criterion is one explanation for the patchy nature of permafrost in Eastern Siberia where topographical differences are large /Ahonen, 2001/.

2.4.3 Soils and vegetation

Most mires are distributed in close connection with areas of frozen ground /Zeyou, 1993/. Peat even though also being an insulator – as well as snow is – is frequently found to be a conservator of permafrost and in the fringing areas of permafrost the sole responsible parameter for sporadic permafrost /Brown, 1974; Williams and Smith, 1989/. During the dry summer months, the peat has very low thermal conductivity (cf section 2.4.4) and therefore stores the permafrost within.

In Scandinavia and Finland it seems to be a rule that large trees do not grow on permafrost ground /Seppälä, 1997/ but this is untrue both in North America and Siberia, where conifers have settled on permafrost.

The degrading effect on permafrost from climate warming may be balanced by uptake of a more productive vegetation and support a northern forest migration /Kolchugina and Vinson, 1993/. The development of a forest canopy may impede an accumulation of snow on the ground /Golding and Swanson, 1986/, and inhibit the solar radiation during warmer periods /Brown, 1974/. Hence the lack of snow coverage and colder summer conditions could decrease the mean annual ground temperature.

2.4.4 Thermal properties of the ground

Thermal conductivity

The thermal conductivity describes the ability of a material to transport thermal energy (e.g. heat). Thermal conductivity depends on a number of variables, such as ground water content, porosity, mineralogy etc. However, except for the uppermost part of the subsurface variations of groundwater content are small and their effect on thermal conductivity negligible /e.g. Gold and Lachenbruch, 1973/. Further for most ground materials the thermal conductivity is dependent on the surrounding temperature in that, the thermal conductivity decreases versus an increasing temperature.

For a dry material the heat conduction is controlled by the heat flow within the minerals and hence thermal conductivity for a dry soil depends on the ground matter and the contact areas within the material /Sundberg, 1988/. /Allen et al, 1988/ reported on a strong observed correlation between the lithology and the permafrost thickness. They concluded that this correlation is best described by the differences in thermal conductivity of different lithological materials.

Table 2-2. Thermal conductivity of some constituents in soils /based on Williams and Smith, 1989/.

Constituent	Mean value [W/(m °C)]
Quartz	8.80
Clay minerals	2.92
Organic matter	0.25

As illustrated in Table 2-2 the mineralogy of the ground is of immense importance to the heat conduction. A quartz rich ground conducts more thermal energy than do an organic soil.

The porosity of a material defines the portion that is available for gases and/or liquids. As such, the porosity has a major influence on the hydraulic- as well as on the thermal properties of the ground. The porosity yields the amount and size of contact areas, therefore the porosity is regularly viewed as the governing variable for heat conduction. The most common soils have porosity values approximately equal to 40 per cent, such soils have in dry conditions a thermal conductivity of approximately 0.3 [W/(m °C)] /Sundberg, 1988/. A low porosity, crystalline rock of similar mineralogy has a much higher thermal conductivity, in the range of 2.5 to 3.5 [W/(m °C)].

The low conduction in a dry and high porosity ground is mainly due to the low thermal conductivity of air. A fully saturated sample would have a much higher thermal conductivity than the same dry sample. Equally a sample would if fully saturated and frozen (ice) have a thermal conductivity approximately four times higher than that of an unfrozen sample /Sundberg, 1988/ (see Table 2-3).

Table 2-3. Thermal conductivity of void constituents in ground /based on Williams and Smith, 1989/.

Constituent	Mean value [W/(m °C)]
Ice (0°C)	2.24
Liquid water (0°C)	0.56
Air	0.025

An increasing ground water content yields an approximately logarithmic increase in thermal conductivity /Sundberg, 1988/ (see Figure 2-1).

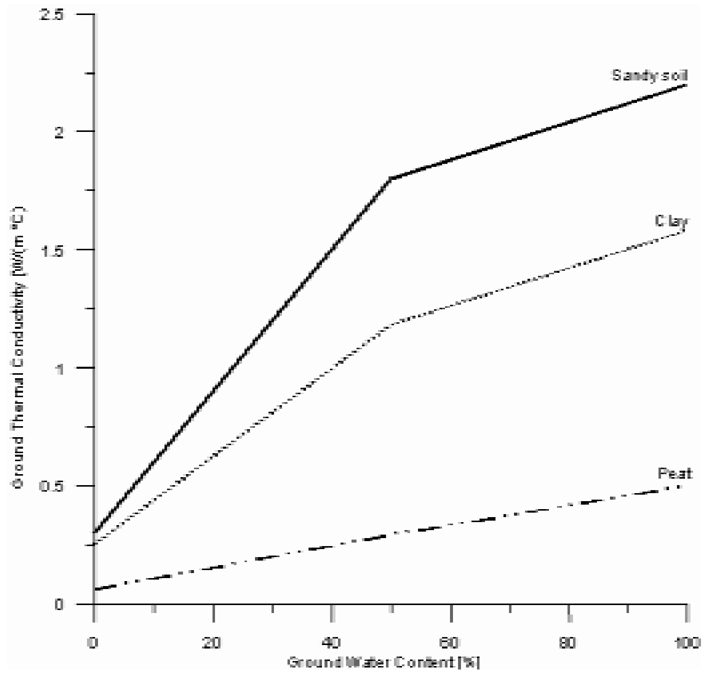

Figure 2-1. *Thermal conductivity as a function of ground water content /based on Williams and Smith, 1989/.*

In Table 2-4 average values of thermal conductivity values for some typical bedrocks from the Fennoscandian shield are presented.

Table 2-4. Example of thermal conductivity of typical Swedish bedrocks (based on material compiled from /Sundberg, 1988/ and /Patel et al, 1997/).

Rock type	Mean value [W/(m °C)]
Greenstone	2.58
Granodiorite	3.34
Granite	3.47

Thermal capacity

The thermal capacity (C) is the ability to store thermal energy (e.g. heat). The most common minerals in rocks show an increase in thermal capacity versus an increasing temperature. The specific heat (c_p) is often used instead of thermal capacity; the specific heat represents the thermal capacity per a unit mass at a constant ambient pressure.

Table 2-5. Example of thermal capacity of typical Swedish bedrocks (based on material compiled from /Sundberg, 1988/ and /Patel et al, 1997/).

Rock type	Mean value [J /(kg °C)]
Greenstone	775
Granodiorite	760
Granite	740

2.5 Heat flow

Governed by the hot core of the Earth and the radioactive decay of elements in crustal rocks, there is a heat flow towards the Earth surface. This heat flow varies from place to place. /Carslaw and Jaeger, 1959/ concluded that there is no systematic variation between different regions. However, later investigations /e.g. Pasquale et al, 1990, 1991; Kukkonen, 1993/ have in theory been able to separate the effect of mantle heat flow from the radioactive heat generation. In doing so, they all conclude that the mantle heat flow shows a systematic behaviour with the smallest heat flux from beneath the centre of old bedrock shields.

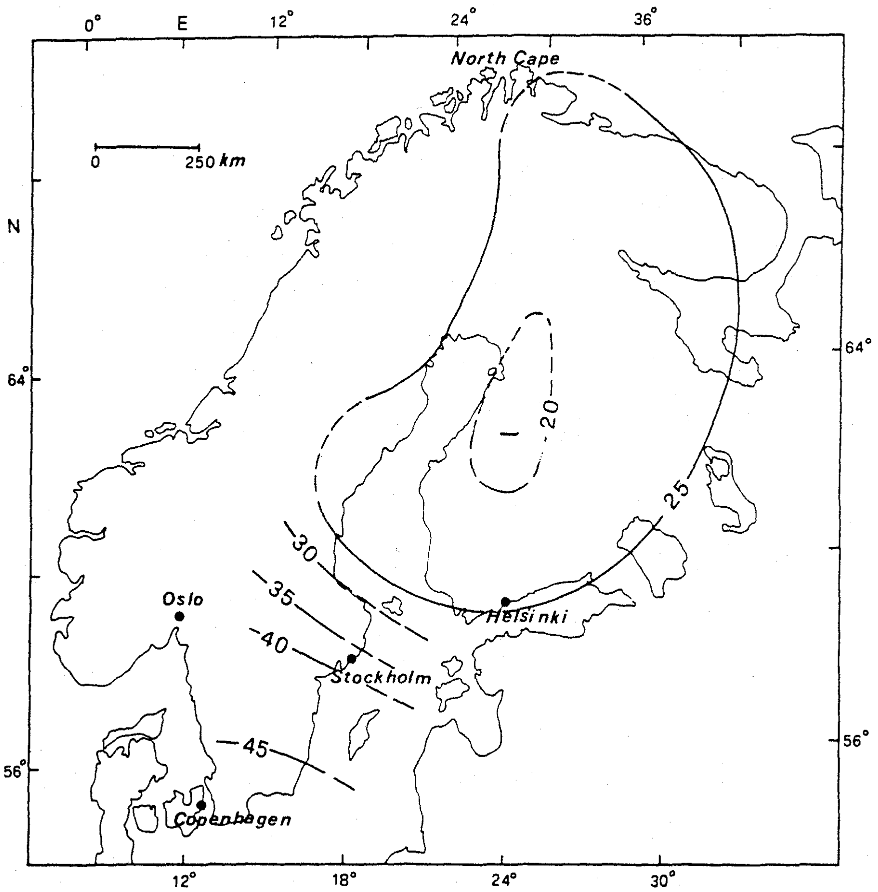

Figure 2-2. *A contoured illustration of the mantle heat flow [mW/m^2] in Fennoscandia /from Pasquale et al, 1991/.*

The geothermal gradient is an expression of the internal heat flow at a location. As such it is one of the most used thermal properties in many modelling investigations. However, the geothermal gradient is one part of the conductive heat flow equation:

$$Q_G = -K \, \partial T / \partial z$$

where

Q_G = Conductive heat flux

K = Coefficient of thermal conductivity

$\partial T / \partial z$ = Thermal gradient (the geothermal gradient).

As a part of the conductive heat flow equation the geothermal gradient could not be viewed as stationary and constant. Many investigations /e.g. Demezhko and Shchapov, 2001; Isaksen et al, 2001/ have illustrated the transient nature of the geothermal gradient, even in deep boreholes. However, after accounting for these climatic signature influences, the geothermal gradient is used as a proxy for heat flow calculations.

If the observed geothermal gradients in a region far from any thermal activity (e.g. vulcanism) are considered in relationship with the thermal conductivity of the bedrock, there will be no large variations in the heat flow. Compilations of global geothermal gradients conclude that the gradient value varies between 6°C per kilometre and 50°C per kilometre on continents far from any volcanic activity /e.g. Carslaw and Jaeger, 1959; Brown, 1974; Demezhko and Shchapov, 2001/, whereas measurements in the ocean floor yield a value approximately 40°C per kilometre. In Sweden, the mean thermal gradient is approximately 15°C per kilometre of depth /e.g. Sundberg, 1991; Hurtig et al, 1992/ in the Precambrian basement. In younger bedrocks, such as in Scania and on Gotland, higher gradients are to be expected. Reported values of gradients from these latter regions are between 25°C and 43°C per kilometre, consequently in Sweden, the thermal gradient can be found in the same scatter as globally observed.

Compilations of heat flow value for Fennoscandia is presented in e.g. /Hurtig et al, 1992/ cf Figure 2-3:

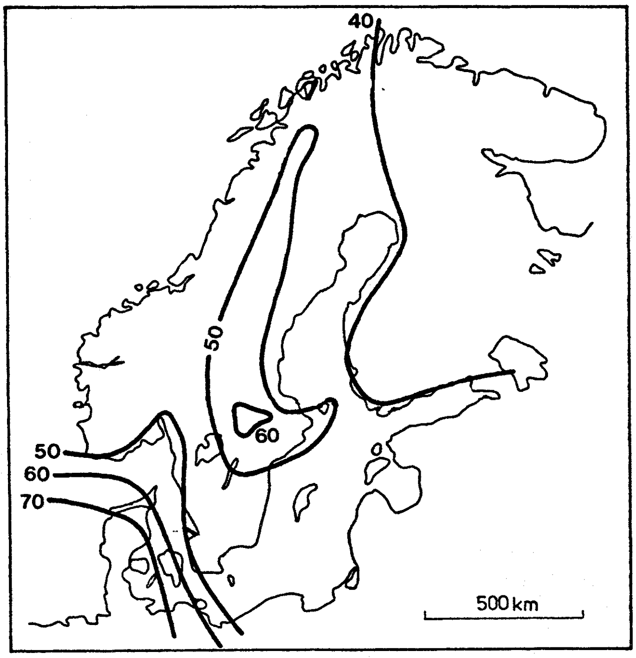

Figure 2-3. *Geothermal surface heat flux [mW/m^2] in Fennoscandia. In the original figure data from /Muir Wood, 1993/ is additionally presented. Muir Wood's data corresponds roughly with the geographical distribution of mantle heat flux (cf Figure 2-2) /from Eliasson and Lundqvist, 1994/.*

3 Permafrost distribution

Permafrost occurrence is reported from both Polar Regions of the Earth. Approximately 25 per cent of the total continental land area at the both hemispheres, is occupied by permafrost with roughly one third in Antarctica on the Southern Hemisphere (Shi in /French, 1996/). About 5 per cent of these 25 per cent are subglacial permafrost on Antarctica and Greenland.

Because of basal melting beneath the Antarctic ice sheet, permafrost occupies less than 25 per cent of Southern Circumpolar Region (that is latitudes ≥ 50°S) /Bockheim, 1995/. However, permafrost is abundant in Alaska, the northern parts of Canada and Russia, and in parts of China on the Northern Hemisphere /French, 1996/. Greenland however, due to the coverage of the Greenland ice sheet, experiences similar conditions as Antarctica, and both continuous and discontinuous permafrost are reported along the coast of Southern and Southwestern Greenland /Mai and Thomsen, 1993/.

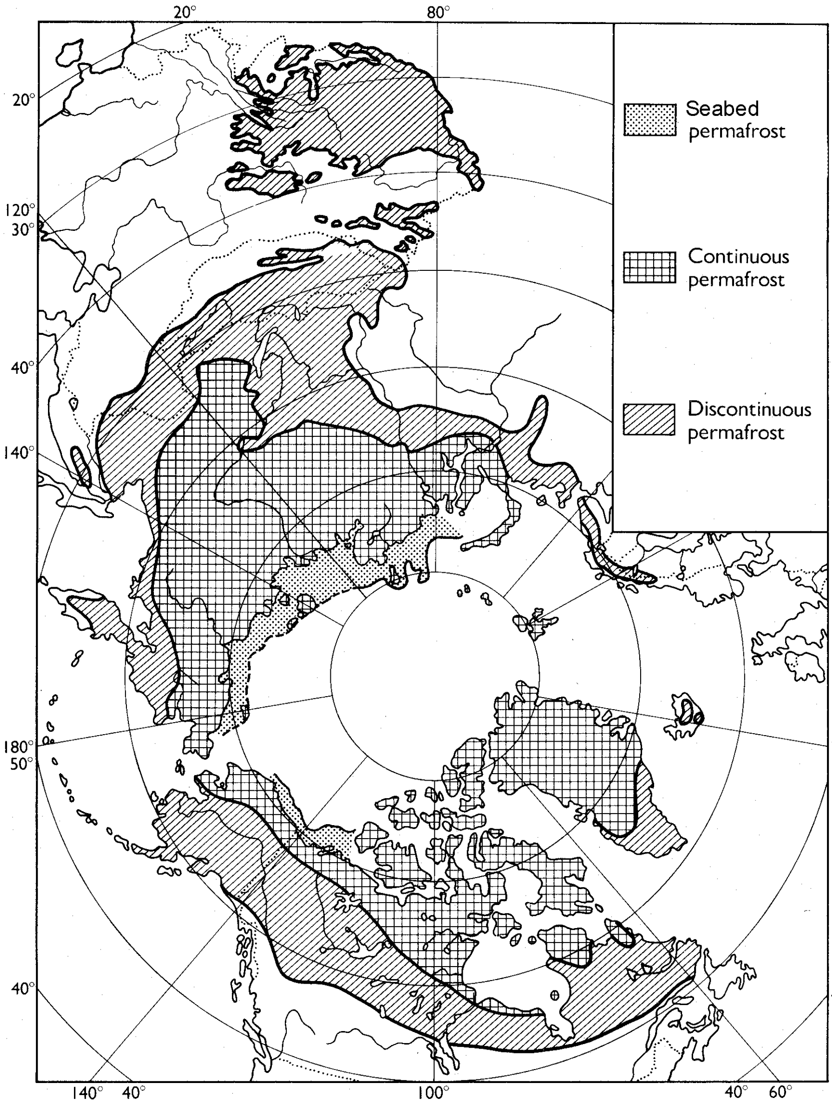

Figure 3-1. *Permafrost distribution in the Northern Hemisphere /from Washburn, 1979/.*

Permafrost is further frequently observed within high altitudes and mountain terrain. In North America mountain permafrost occurs along both the continental margins /Schmidlin, 1988; Harris, 1993/, extending all the way down to Mexico where it is reported to occur above 4,500 metres /Péwé, 1983/. Extensive mountain permafrost is also common both in high altitude Europe /Washburn, 1979; King and Åkerman, 1993; Sollid et al, 2000/ and Asia /Guoqing, 1993/.

It is common practice that the permafrost zones relate to a specific ground surface temperature and the sporadic permafrost occurs in regions where the temperature is close to 0°C and where localised favourable conditions, such as a peat cover enhance/conserve the permafrost /Williams and Smith, 1989/. The lower temperature limit of continuous permafrost is frequently accompanying the –8°C mean annual air temperature (MAAT) isotherm in North America /Brown, 1974/ and the –7°C isotherm in Russia /Åkerman, 1980/. These temperatures correspond roughly with a ground surface temperature of approximately –5°C /Brown, 1974/.

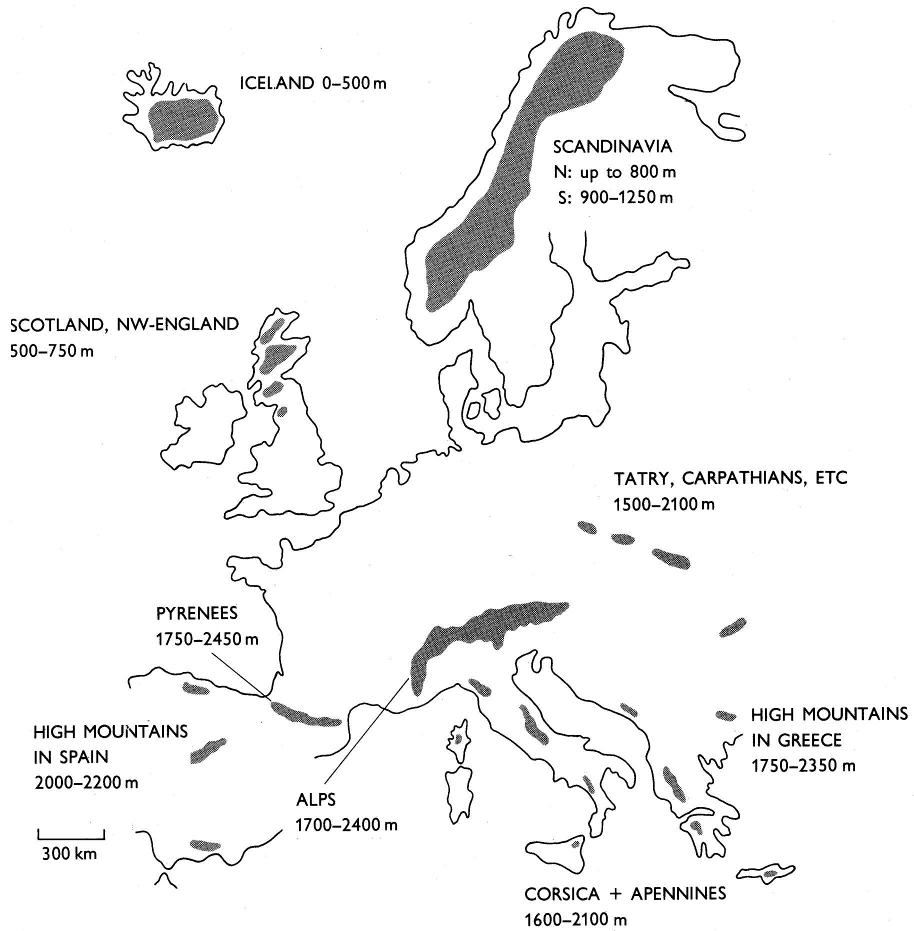

Figure 3-2. *Mountainous areas in Europe from where periglacial features have been reported and therefore imply potential permafrost regions. Numbers represent heights of the lower altitude boundary of periglacial features /from Washburn, 1979/.*

Permafrost distribution has both an areal component and a thickness or depth component. According to /French, 1996/, as a rule-of-thumb the stationary thickness (depth, d_{perm}) of permafrost can be estimated by:

$$d_{perm} = \frac{MAAT}{\partial T / \partial z}$$

where

d_{perm} = Permafrost depth

MAAT = Mean annual air temperature

$\partial T / \partial z$ = Geothermal gradient.

This rule-of-thumb has, however, been found only to give a very rough estimate of the permafrost depth. More processes than those described by this rule-of-thumb are involved in the evolution of permafrost (cf section 2). However, the rule-of-thumb do describe a simplistic view of the conductive heat flow equation and as such it implies that a cold climate is accompanied by deep permafrost. However, /Lunardini, 1993/ concluded that permafrost at greater depth (approximately below 600 metres) grows on the time scale of several ice ages, that is many 100,000 of years (cf section 3.1.4 and 3.1.5). Additionally even surficial permafrost grows at a relatively slow rate /Lunardini, 1993; Delisle, 1998/ and therefore a presently cold climate is not always equivalent to deep permafrost. This since the time necessary for a great permafrost thickness to develop far exceeds the time available on the natural scale of climate variation.

3.1 Continental permafrost

Continental permafrost is common on the Northern Hemisphere; continuous, as well as discontinuous and mountain permafrost. In most of these regions a well-documented correlation between the mean annual air temperatures and the permafrost distribution is found.

3.1.1 Scandinavia

In the high altitude fringe of Northern Europe (e.g. the Kebnekaise massif in Northern Sweden and the Jotunheimen area in Southern Norway) extensive regions of permafrost occur (cf Figure 3-2).

The permafrost in the area of Tarfala (part of the Kebnekaise massif) at an altitude above 1,500 metres was found to be more than 100 metres thick /King, 1984/. King defined this high altitude permafrost area to be part of the continuous permafrost region. Further the permafrost that exists below, between the altitudes of 1,200 to 1,500 metres, was defined as discontinuous. /Jeckel, 1988/ found ground temperatures in shaded areas or on north slopes that indicated permafrost at the Abisko area (at approximately 900 m a s l) in Northern Sweden. For the north facing slopes the permafrost might even extend downwards lower altitudes and Jeckel assumed the lower altitude limit for the discontinuous permafrost in the Abisko area to be around 800 metres.

In general, the active layer in the Tarfala area has a thickness of 1.3 metres and at the altitude of 1,500 metres the mean annual air temperature is approximately –6°C with the ground surface temperature about 2 degrees warmer /King, 1984/.

In the Jotunheimen area the permafrost thickness is between 100 and 200 metres at 2,200 metres height. Here the ground surface temperature is approximately –6°C as a yearly average. The active layer is somewhat thinner compared with the active layer found in the Tarfala area. In Jotunheimen sporadic permafrost is found at 1,000 metres height in peat bogs /King, 1984/.

Based on above described findings King has created a vertical zonation of permafrost in the central parts of the mountainous Scandinavia (see Figure 3-3). In this zonation permafrost is distributed by temperature with the continuous permafrost corresponding to the –6°C isotherm and the lower limit of permafrost potential along the –1.5°C isotherm. These figures correspond approximately to other figures of temperature zonation used /e.g. Brown, 1974; French, 1996/.

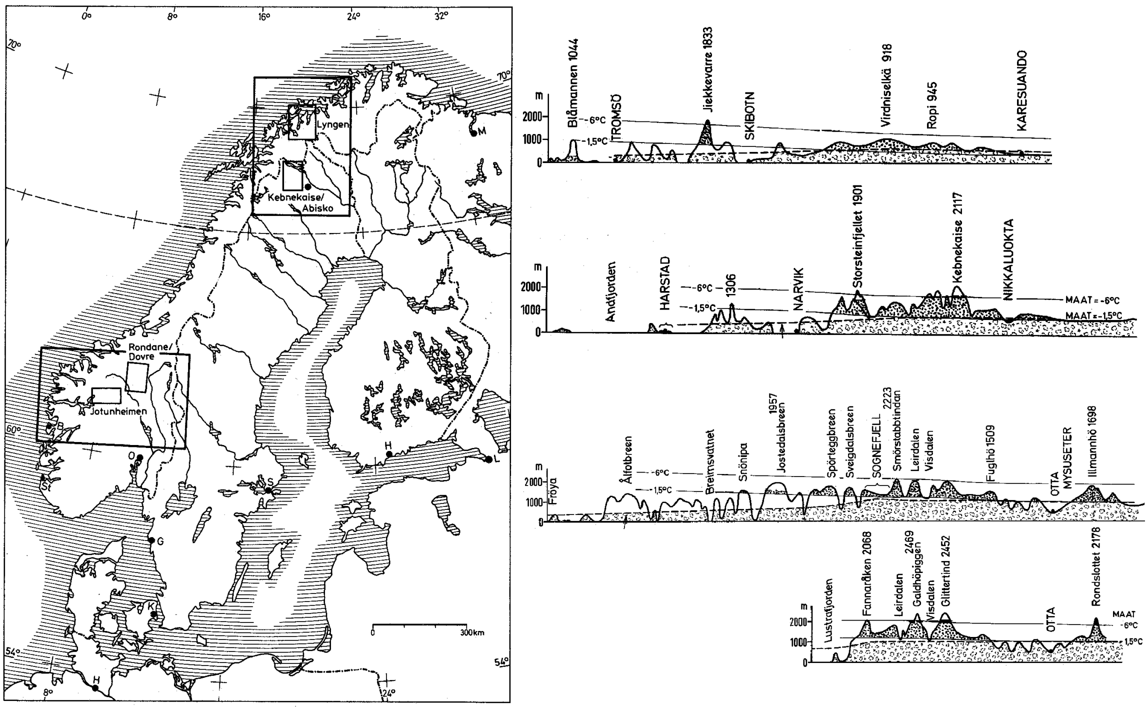

__Figure 3-3.__ Cross-sections through the Scandinavian Mountains. The upper two cross-sections are from the northern parts of Lapland (see boxes in the side map) and the lower two are from Southern Norway. The –1.5 °C and –6 °C MAAT isotherm lines indicate the permafrost potential. The lower dashed line in the cross-sections illustrates the tree line /from King, 1984/.

3.1.2 Finland

Compared to the northern parts of Norway and Sweden, Northern Finland is flatter and lower in altitudes, with only few summits exceeding 1000 metres in the Northwest of Finland /King and Seppälä, 1987/.

Deep frost in Finnish Lapland has been attributed mainly to thin snow covers /Seppälä, 1990, 1994/ on mire grounds. A development of a continuous permafrost reaching significant depths in Finland would require a considerable decrease, in both temperature and precipitation /Ahonen, 2001/. However, /King and Seppälä, 1987, 1988/ concluded that the permafrost in Northern Finland extends southerly beyond the Palsa Region (cf Figure 3-4) and that frozen ground is widespread on barren mountainous areas above the tree line.

Mountainous permafrost on north slopes is present down to approximately 600 metres altitude in Northern Finland /Jeckel, 1988/ and the potential for extensive permafrost should exist in areas of approximately 750 metres altitude or higher. However, /King and Seppälä, 1987, 1988/ stated the fact that these kinds of altitudes are rare in Finland.

/King and Seppälä, 1987; Seppälä, 1997/, among others, have reported on permafrost conditions down to a depth (thickness) of 60 metres in the Finnish Lapland. /Kukkonen and Šafanda, 2001/ reported on permafrost of this depth as far to the south as Ylläs (Southwest part of the displayed region of Finland in Figure 3-4).

Observations have indicated that in many permafrost areas in Northern Finland, the active layer is deeper than would be expected due to present temperatures, therefore it is discussed and regularly implied by the investigators that some of the observed permafrost is relict /e.g. King and Seppälä, 1987/.

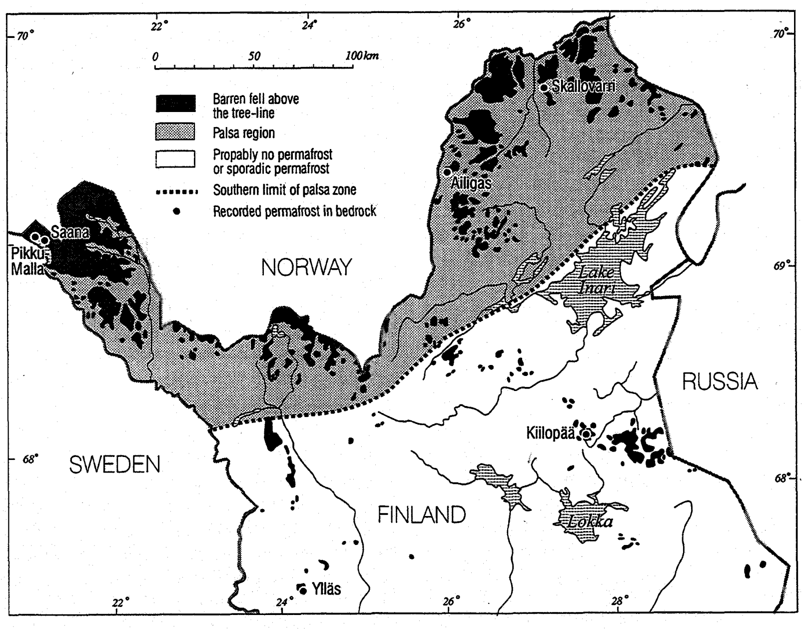

Figure 3-4. *Areas where a potential for permafrost development exists in Northern Finland. Within the Palsa region, permafrost is frequently found in peat on mires. On the mountainous areas permafrost is found dependent on ground matter, topographical features, and elevation /from Seppälä, 1997/.*

3.1.3 Spitsbergen (Svalbard)

The climatic situation in the region of Svalbard is suitable for permafrost. The permafrost on Spitsbergen (part of Svalbard archipelago) is continuous, with some open taliks of different origins and some sub-glacial groundwater recharge areas at the base of some temperate sub-polar glaciers. This continuous permafrost has a thickness that varies between approximately 100 and 400 metres /Åkerman, 1980; Haldorsen and Heim, 1999; Isaksen et al, 2001/.

The active layer on Spitsbergen is frequently found to be as much as two metres thick /Repelewska-Pekalowa and Gluza, 1988/

Spitsbergen has lots to offer as a contemporary analogy to the environment of Western Europe during the late stages of the Weichselian glaciation /Åkerman, 1987/. Despite the high latitudes of Svalbard there are great climatic similarities between present-day Spitsbergen and the Western Europe some 10,000 to 20,000 years ago.

3.1.4 Alaska (USA) and Canada

The broad outline of permafrost in North America is well documented /French, 1996/. In Canada nearly 50 per cent is affected by perennial frozen ground and in Alaska, as much as 80 per cent of the land is underlain by permafrost.

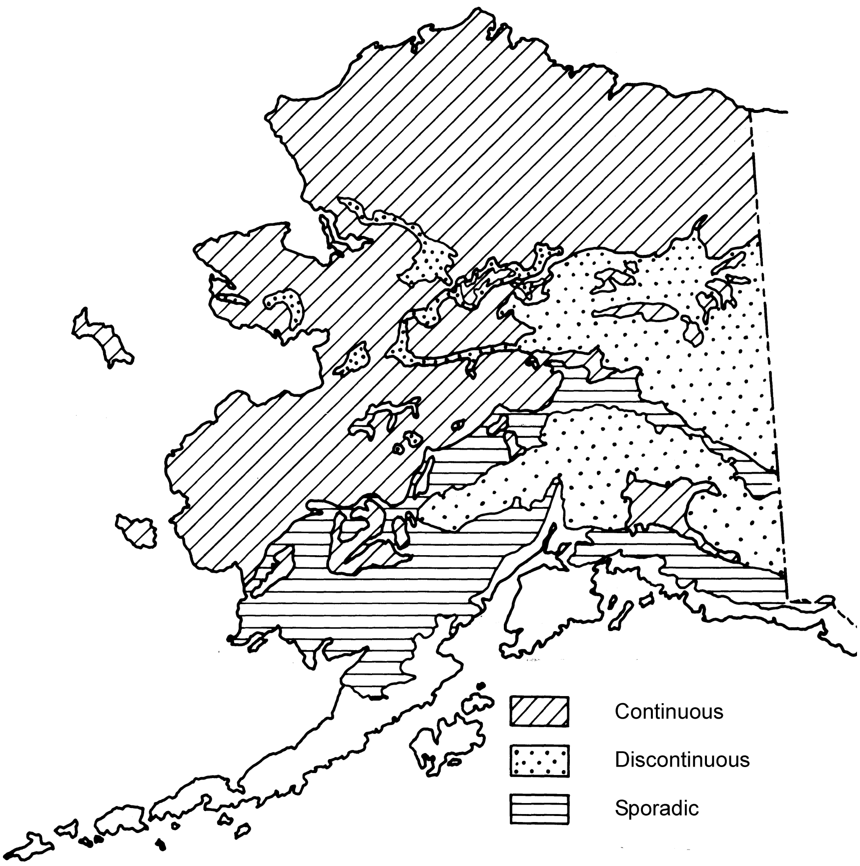

Figure 3-5. Permafrost distribution in Alaska /from Carlson, 1974/.

In Alaska, a continuous belt of permafrost dominates the northern and much of the western parts. In the north of Alaska the known thickness in this continuous zone is more than 500 metres. The deepest known permafrost in Alaska is found in Prudhoe Bay, with a permafrost depth of 610 metres /French, 1996/.

Most of the Northwestern Territories in Canada is underlain by thick and continuous permafrost. In Canada as well as in Alaska the southern line of the continuous permafrost zone is well correlated with the climate and especially the temperature. Field observations in Canada indicate that a temperature isotherm of –8 to –6°C coincide with this southern boundary of continuous permafrost /Brown, 1974/.

In the continuous permafrost zone of North America permafrost occurs everywhere except for underneath large water bodies and within newly deposited sediments in among other river delta plains and alluvial fans /French, 1996/.

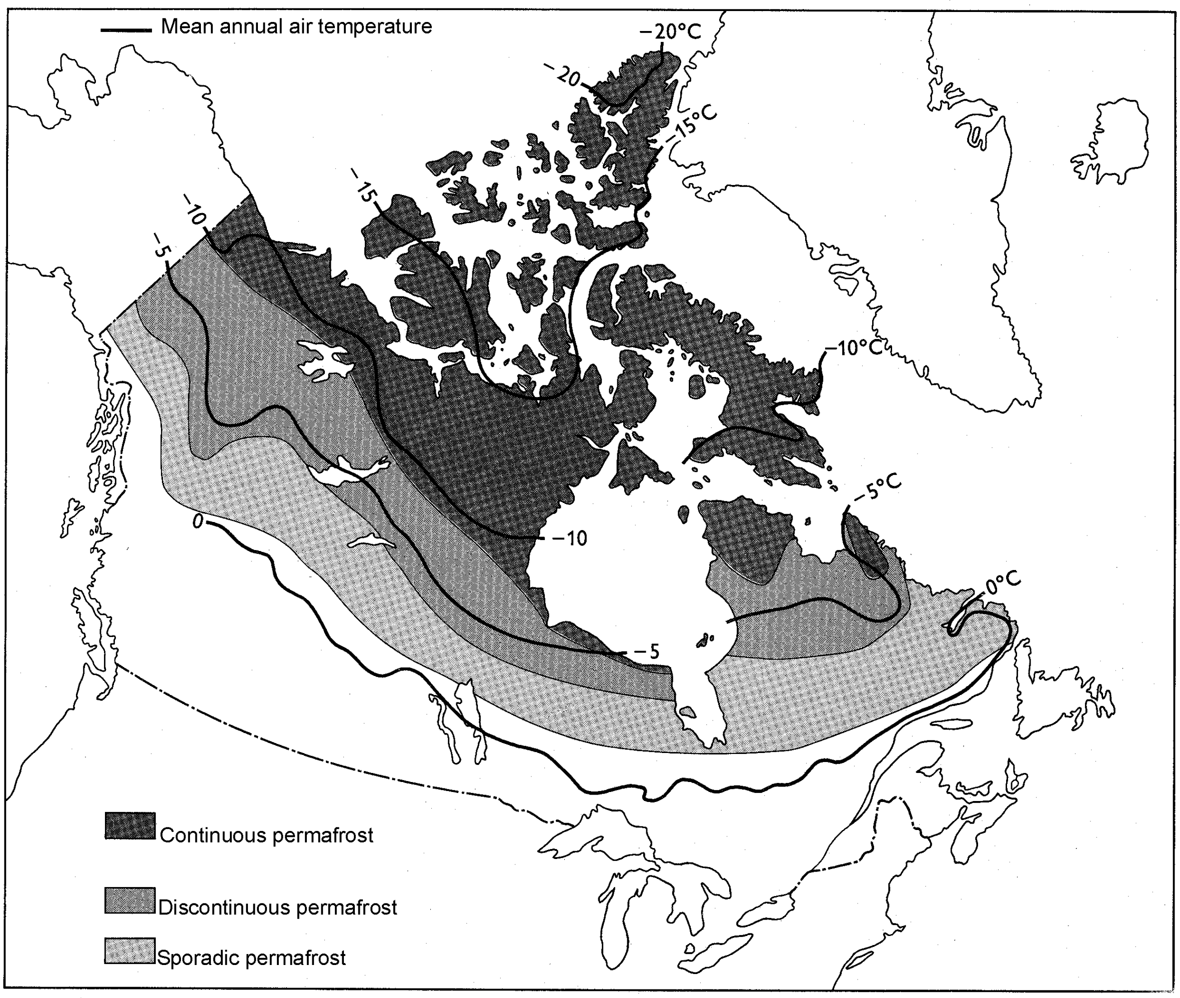

Figure 3-6. *Permafrost zones in Canada; no mountain permafrost is accounted for /from Washburn, 1979/.*

The thickness of permafrost in Canada ranges from just a couple of metres in the sporadic permafrost zone in the south to the much deeper permafrost in the Northern Archipelago and Mainland. Thickness values of more than 300 metres are reported in the maritime environment of the northern part of Hudson Bay /Washburn, 1979/. On specific islands in the Arctic Archipelago thickness values of 400 to 600 metres are common /e.g. Judge, 1973/.

The areas of North America containing the deepest permafrost are all regions that were ice-free during most of the time of the last glacial cycle.

3.1.5 Russia and China

Permafrost underlies approximately 50 to 70 per cent of Russia and 20 per cent of China /French, 1996; Fotiev, 1997/. Except for the westernmost part of the North Coast of Russia, continuous permafrost with a maximum thickness of approximately one kilometre in the central Siberia occupies all land areas /Yershov, 1998; Ahonen, 2001/. This continuous permafrost region is separated only by taliks /Yershov, 1998; Ahonen, 2001/ under the great rivers, i.e. Lena and Yenisey. The region of the thickest permafrost corresponds to the Anabar massif, a crystalline rock massif of Archean age. This rock massif's lower thermal conductivity compared to the surrounding younger sedimentary rock's thermal conductivity values yields a lower geothermal gradient and the frozen fringe has been able to migrate deeper /Yershov, 1998/.

The deepest known permafrost occurs in the central part of Siberia, with thickness values of up to 1,500 metres /Fotiev, 1997/. As deep permafrost demands extensive time in order to develop (cf section 3.3) these central regions of deep permafrost correspond to previously unglaciated regions.

3.2 Sub-sea permafrost

The definition of permafrost often results in that areas are defined as sub-sea permafrost regions, even though no ice is present /French, 1996/.

Deep-drilling for gas and oil exploration /Gascoyne, 2000/ and seismic and temperature observations /Allen et al, 1988/ in offshore regions have resulted in the findings of permafrost beneath newer sediments at the sea bottom.

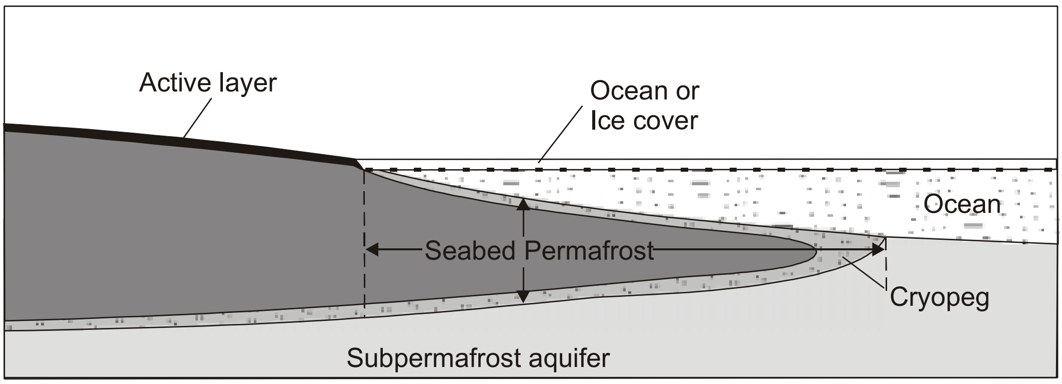

Figure 3-7. *Sub-sea permafrost definitions /from van Everdingen, 1990/.*

Areas of sub-sea permafrost that contain ice are in general believed to be the relict result of permafrost evolution when these seabeds were exposed to a cold climate during low sea level event in Pleistocene /Allen et al, 1988/.

Even though seawaters in permafrost regions are, in general, below 0°C the energy balance is positive in relationship to the historic surface temperature and hence a degradation of the sub-sea permafrost occur world-wide /Gascoyne, 2000/. However, this process of permafrost degradation and also that of aggradation of sub-sea permafrost are complex. Dyke /in Gascoyne, 2000/ has described the transient nature of these two processes. Summarised, permafrost degradation starts when the sea level rises and the area experiences shoreline erosion. The permafrost degradation process is frequently enhanced by a discharge of warmer water from river mouths. However, when water levels rise the thawing ceases and the seabed, in general, once again develops permafrost, controlled by the deeper and colder levels.

All frozen seabeds are not by necessity relict. A rich discharge of fresh waters along the seabeds in the supercooled Arctic waters could keep a low salinity in the bottom sediments and therefore the sediments might as a consequence freeze.

3.2.1 Shoreline effects

The shoreline is an exposed region; tides, waves, erosion, seawater depth, land rise, and a harsher climate /Allard et al, 1988; Dyke, 1988/ all contribute to different characteristics of permafrost present along the shoreline. In a region with mean annual air temperatures of between –5 and –6°C /Allard et al, 1988/ found that permafrost could develop in areas of the shore that were above sea-level for more than 10 per cent of the year. These permafrost 'mounds' had a thin layer of permafrost of about 2 to 6 metres.

In shore parts, with ground surface above the extreme shore (tide) level permafrost develop however in a thin layer, generally bounded by the sea water level in the ground /Allard et al, 1988/. The ground temperature distribution in these areas is governed by the climate, but also the movement of saline groundwater can cause the permafrost to be sporadic and profoundly effect the total permafrost thickness that can develop.

One specific shoreline effect is the isostatic uplift that many previously glaciated regions experience today. Such uplift will in cold environments yield an aggradation of permafrost.

3.3 Permafrost reconstruction

Permafrost is mainly a thermal feature and does as such respond to climatic changes; however the response to climatic warming and cooling is slow /Osterkamp, 1984; Dawson, 1992/. Based on the fact that ice functions as an insulator for escaping geothermal heat; in theory, the existence of large ice sheets could have impeded and slowed down the growth of permafrost. However, thermal isolation from ice sheets does not fit some of the present distributions of permafrost and/or permafrost proxies, therefore it must be considered possible that the great ice sheets advanced over regions already underlain by permafrost. Today methods for reconstructing permafrost distribution and associated climatic situations are available. In for example /Bradley, 1985/ and /Isarin, 1997/ reviews on the ability and validity of such methods are presented.

South of the Laurentide ice sheet on the North American continent only a relatively narrow belt of permafrost seems to have occurred /Wayne and Guthrie, 1993; French, 1996/. /French, 1996/ argued that this partly may be due to the much lower latitudes of this ice sheet compared to the Eurasian ice sheet described below. North America was during the last glacial cycle exposed to two ice sheets, the Laurentide to the east and the Cordillian to the west. During their maximum extension the ice sheets were to some extent merged. The two ice bodies created atmospheric conditions yielding dry and cold climate in the interior of the North American Continent and also in the exposed land of Alaska, Northwestern Canada, and the water free continental shelf of Beaufort Sea, Bering Strait, and its soundings. These climatic conditions were suitable for an extensive permafrost development, especially in the northwestern parts of North America /Dawson, 1992/.

The beginning of the last stages of the Weichselian glaciation appears to have been the coldest period with extensive permafrost activity within Eurasia /e.g Velichko, 1984; Dawson, 1992/. The distribution of sea- and land ice, and permafrost is closely linked with the local climate. To the east of the Weichselian ice sheet very dry air flowing down from the ice sheet invaded the unglaciated areas /Velichko, 1984/. This in association with the southward displacement of the polar/oceanic atmospheric front, and possibly air jet streams flowing along the margin of the Weichselian ice sheet both from the north and south toward east caused a stationary air mass over Eastern Asia that produced a cold and arid environment. This environment intensified the development of permafrost west to southwestward /Velichko, 1984; Yershov, 1998/.

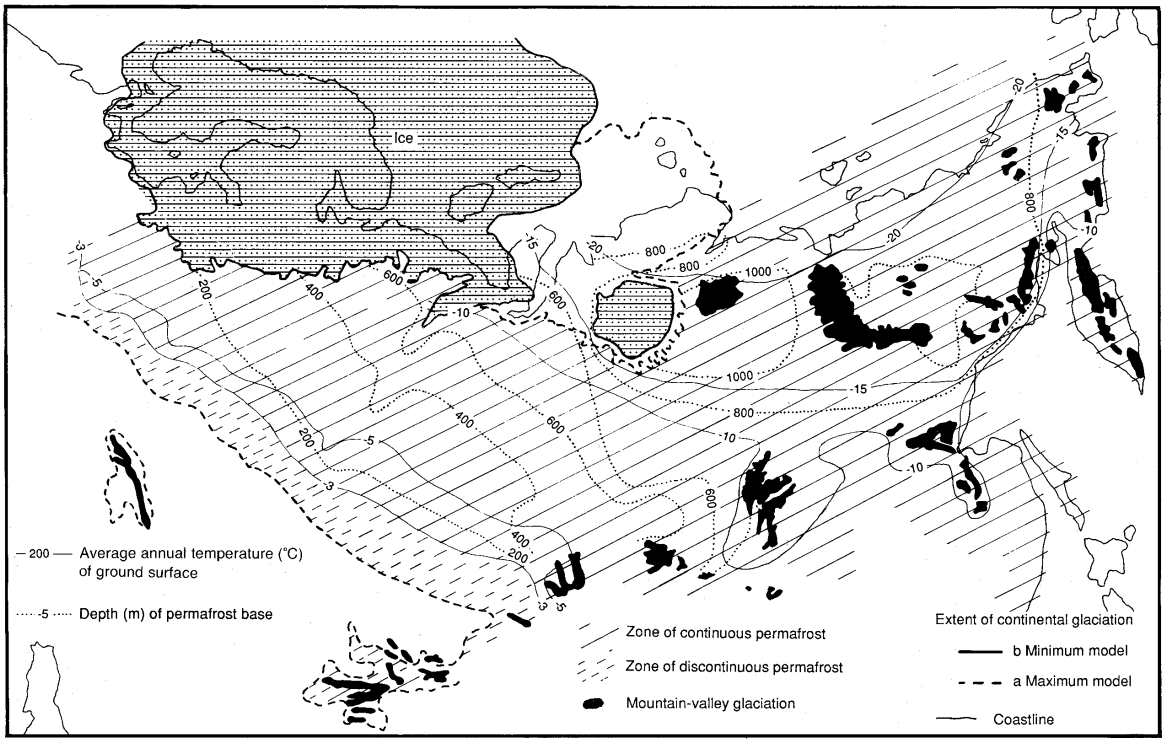

Figure 3-8. Map of permafrost distribution to the east of the Eurasian ice sheet during the Weichselian glaciation /from Dawson, 1992/.

In addition to the climatic conditions previously stated, an ice cap over the Alps caused the mid-latitude atmospheric jet stream to be displaced further southward affecting the Mediterranean with large precipitation rates /Dawson, 1992/. However, north of the Alps an easterly, cold and dry wind from the southern and eastern parts of the Weichselian ice sheet supported a southerly growth of permafrost. On the Iberian Peninsula the temperature dropped considerably during the glacial periods in Eurasia and this part of Europe, merged in-between the Mediterranean and the Atlantic Sea, had an arid climate with sporadic permafrost /European Commission, 1997/.

In climatic reconstructions the precipitation amount is hard to assess /French, 1996/. In general, a colder climate and the blocking of westerly air streams are believed to have caused an overall decrease in precipitation over Europe during the Weichselian glaciation. In the uplands of Western Britain, Manley /in French, 1996/ suggested a precipitation value of approximately 80 per cent of the present day (1959) value. If this historic value showed similar relationship with the rest of Europe then as today much less precipitation would have fallen in Central and Eastern Europe /French, 1996/. Such values would have resulted in the presence of a desert environment, similar to the present day Arctic desserts, with less than 200 mm precipitation per year in Northeast Europe. This kind of value is comparative with many high Arctic regions of today.

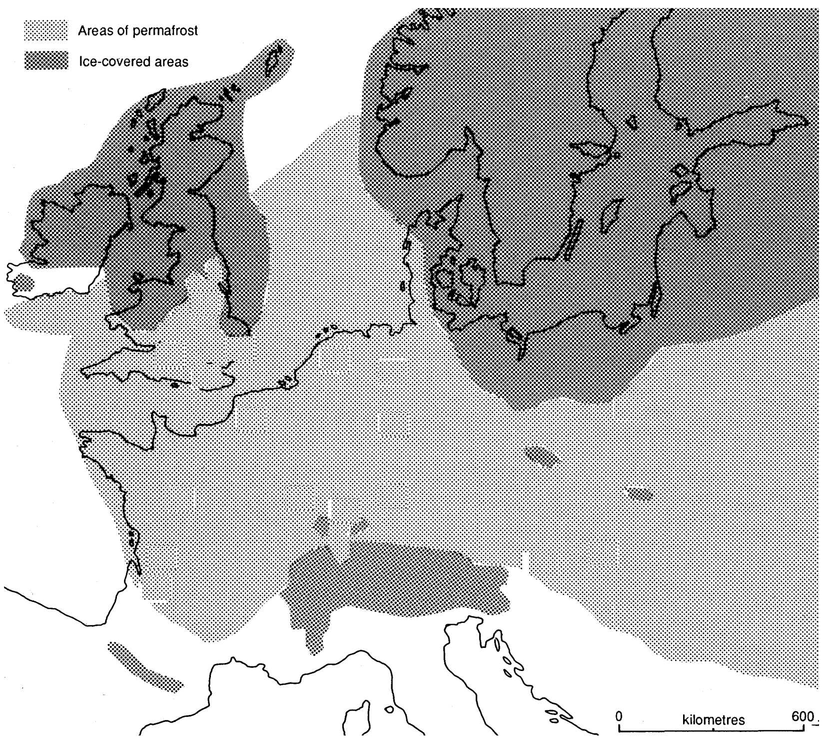

Figure 3-9. *Reconstructed permafrost distribution in Europe during the Weichselian maximum /from Dawson, 1992/.*

/Isarin, 1997/ compiled information on permafrost proxies in Europe, which based on geochronological control measures have been dated to the Younger Dryas. In Fennoscandia ice-wedge polygons and solitary wedge casts indicate perennial frozen ground. /Black, 1976/ suggested that ice-wedge proxies alone do not conclude on a continuous permafrost but are indicators on locally favourable climate conditions. However, ice-wedge polygons are abundant in Southern Sweden, particular in Scania (Skåne) and the central highlands of Southern Sweden. On the Swedish West Coast (e.g. Halland) proxies such as solitary wedge casts are reported. Based on these proxies /Isarin, 1997/ concluded that the ice-free land areas of Younger Dryas Fennoscandia were exposed to continuous permafrost.

/Isarin, 1997/ further suggested a temperature distribution with a mean annual air temperature, at the sea surface, of –8°C corresponding to the change between continuous and discontinuous permafrost. Isarin stated that these temperature results are largely in accordance with fossil beetle proxies, which according to Isarin indicate a mean air temperature of the coldest month (MTCM) of below –20°C in Southern Sweden and on the British Isles. /Vandenberghe and Pissart, 1993/ also concluded that Central and Western Europe experienced discontinuous permafrost in sea level areas during Younger Dryas. Higher altitude localities may have experienced even colder and dryer conditions.

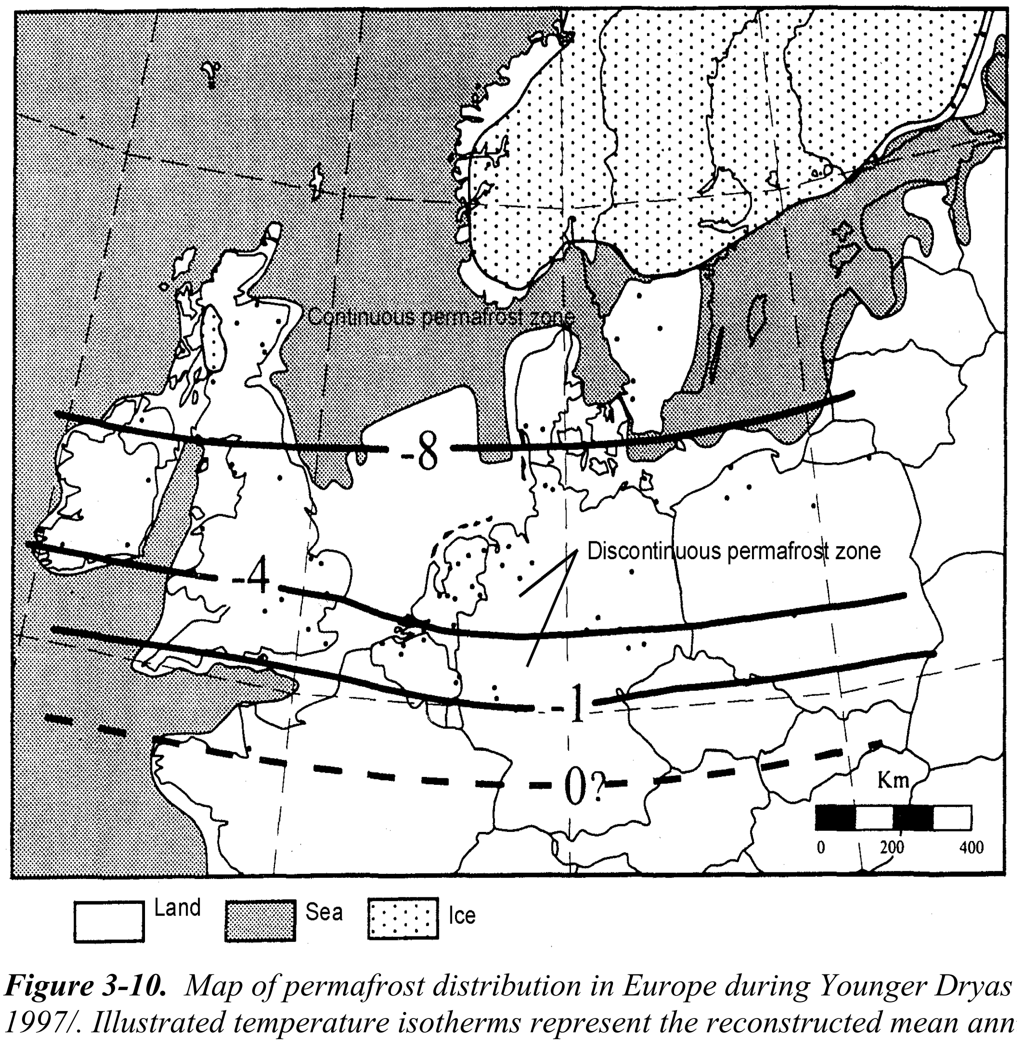

Figure 3-10. *Map of permafrost distribution in Europe during Younger Dryas /from Isarin, 1997/. Illustrated temperature isotherms represent the reconstructed mean annual air temperature isotherm lines.*

Based on reconstruction of vegetation with a zonal method /Savina and Khotinskiy, 1984/ concluded that the total annual precipitation during the Boreal period (9,000–8,000 years BP) was greater in the Northern and Northeastern Russia and less in areas south of latitude 60°N compared with present day precipitation.

Little is known on the permafrost that may have existed in Fennoscandia before and during the last glaciation. Since the entire landmass was covered with ice most proxies have been destroyed. Some evidence may be present in the hydrogeochemistry, but this is not yet fully understood. Numerical model simulations /e.g. Boulton and Payne, 1992/ have suggested that permafrost was preceding the ice coverage in many areas of Fennoscandia. In the mountainous areas where the ice sheet started its growth a permafrost developed early, this permafrost is believed to have survived during the entire glacial cycle /Boulton and Payne, 1992/.

4 Permafrost hydrology

The presence of frozen ground is of immense importance to the hydrology. At a macro-scale perspective the traditional elements (e.g. intensities, magnitudes etc) of the hydrological cycle are all modified in significance. Additionally, the groundwater regime (hydrogeology) is further divided into frozen and unfrozen parts. The presence of a permafrost layer will create a complex hydrology with intense and short peaked surface run-off. Permafrost will further act as a more or less impermeable, confining layer, that inhibits and re-directs recharge and groundwater flow.

The water movement in permafrost ground is not only affected by the presence of soil, water, and air but also of ice present. Further fluid properties change significantly when the temperature of ground water is decreased /Kane and Stein, 1983/. For most purposes of groundwater flow the frost table (i.e. the top of a frozen layer) is viewed as 'impermeable' /Church, 1974; Woo and Steer, 1983/ and consequently frozen ground and especially a permafrost layer is viewed as an aquiclude or aquitard. This perspective yields a group of different aquifers convenient to use in permafrost regions (Sloan and van Everdingen in /McEwen and de Marsily, 1991/):

- Suprapermafrost aquifers
- Intrapermafrost aquifers
- Subpermafrost aquifers.

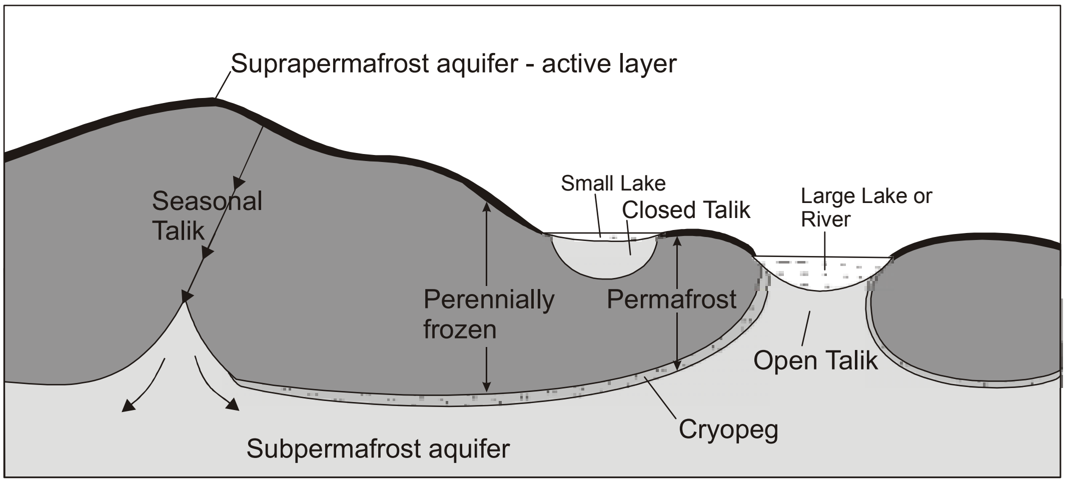

Figure 4-1. *Some permafrost definitions and representation of some typical aquifers in permafrost regions /from van Everdingen, 1990/.*

Suprapermafrost aquifers are situated above the permafrost layer, and basically constitute of large water bodies on the ground surface, of liquid water in the active layer, and of water sources below the active layer but on top of the permafrost.

The *intrapermafrost aquifers* are found in unfrozen zones within a permafrost layer. The intrapermafrost aquifers correspond with the frequent notation of taliks. As an aquifer the talik may function as a conduit between the supra- and subpermafrost aquifers (open taliks), however some may be connected in one direction only (closed taliks) and further some may be isolated. Following /McEwen and de Marsily, 1991/ taliks are divided into:

- Thermal taliks
- Hydrothermal taliks
- Chemical taliks
- Hydrochemical taliks
- Pressure taliks
- Piezochemical taliks.

This grouping of taliks refers to 'generic' conditions or simply the reason for the taliks to stay unfrozen.

The *subpermafrost aquifer* is situated below the permafrost and therefore has water temperatures above 0°C. The extension of subpermafrost aquifers depends on the hydrogeology of the substratum and can therefore be extremely varied.

4.1 The active layer

On top of a permafrost layer, an active layer exists. The thickness of the active layer may be of centimetre to metre scale. In this layer repeatedly freezing and thawing occur during a year cycle. Permafrost as a thermal feature responds slowly on a human time scale to both warming and cooling /Osterkamp, 1984; Dawson, 1992/, however there is evidence on quick responses on the active layer thickness /Kane et al, 1991/. Just like the large-scale permafrost processes, the processes in the active layer are dependent on the local climate and thermal properties at the site /Brown, 1974; Åkerman, 1980; Repelewska-Pekalowa and Gluza, 1988/.

For some time of a year parts of an active layer will be over-saturated with liquid water and for some other time the ground will be frozen and partly impervious for groundwater recharge and flow.

4.1.1 Surface run-off

Despite the great diversity of the permafrost environment some hydrological behaviours are common for them all /Woo, 1990/. For most aspects the surface run-off dominates as the discharge process of received water, which is dominated by precipitation and/or melt of snowbanks and/or glacial melt /Kane and Hinzman, 1988/. The latter is dependent on the proximity to a glacier.

A hydrological basin with late-lying or perennial snow develops a discharge regime, similar to proglacial watersheds /March and Woo, 1981; Lewkowicz and Young, 1991/. In such a basin the input of water is totally dominated by the snowbank melt and the discharge of water is mainly by surface run-off. /Lewkowicz and Young, 1991/ further concluded that the effect of the snowbank melt diminishes with the decreasing snowbank size and hence with the percentage of the basin covered by snow.

The surface run-off is mainly dependent on two processes: streamflow in streams, rivers, etc, and on-surface flow in hillslope regions during specific times of the year, such as the over-saturated spring meltwater fluxes. Further, for some parts of the cold period the surface run-off will perish and returns again with the spring /Woo, 1990/. Of additional importance in cold region hydrology is the presence of lakes, wetland and other open large water bodies /Woo, 1990/. Thaw lakes develop during the spring and may experience a sudden drainage when somewhere an erosional channel is cut through. Such events are mostly found in relation to a path of interconnected ice-wedges /Brewer et al, 1993/.

Streamflow

Streamflow in continuous permafrost regions often follows the flow patterns of nival catchments /Woo et al, 1994/. The main part of the annual run-off follows the snowmelt with additional events, such as: summer rainfalls, snowfalls, and melt of ground ice in the active layer causing sporadic high flows. The permafrost effects the streamflow indirectly through the high groundwater table within the active layer /Dingman, 1973/

In the discontinuous permafrost zone in Canada the average groundwater contribution to stream baseflows generally ranges from 2.0 to 5.0 litres per second and square kilometre (corresponding to 60 mm to 150 mm annually)/Williams and van Everdingen, 1973/. Further to the north in the continuous permafrost zone many streams approach a no flow situation during the winter. This is due to the accumulation of water within icings and the closure of aquifers within the active layer /van Everdingen, 1974, 1990/.

Snow hydrology

The lateral water flow through a saturated slush layer at the base of a snowbank /Colbeck, 1974; Marsh, 1990/ can be viewed as a special kind of surface run-off. Such a meltwater discharge demands that the ground is impermeable. The criterion that the ground is impermeable is, however, mostly unsatisfied /Burt and Williams, 1976; Kane and Stein, 1983; van Everdingen, 1990/ and the meltwater is able to infiltrate into the ground. But, if the ground is fully saturated no more water can infiltrate the ground. Hence this saturated slush layer process is likely to drain the snow mass. An unfrozen and/or unsaturated organic soil layer with comparatively high hydraulic conductivity is surmised to mimic such a hydraulic process /Quinton et al, 2000/.

4.1.2 Infiltration and recharge

Precipitation in the form of rain and meltwater from snow and ice yields water for ground infiltration. Rainwater on a bare ground can infiltrate directly following traditional hydrological rules. However, the water from snowmelt first needs to find its way down to the permeable ground surface. The pressure potential and the gravitation govern infiltration processes. But, equally important is the permeability of the ground matter. The permeability depends on, among other factors, the ground water content. A fully saturated ground sets the upper limit for infiltration. If the ground is frozen, the case of a fully saturated ground is significantly affected towards a lower permeability and therefore a frozen, saturated ground has a very low infiltration capacity.

In permafrost areas the heat flow from ground yields insignificant snowmelt /Woo, 1986/ because the ground temperature is usually below zero as long as the snow cover remains. However, meltwater can reach the base of the snow cover through meltwater flow fingers /March and Woo, 1984/ and such meltwaters could penetrate into the frozen ground below /Woo, 1986/ or possibly be discharged through a slush layer within the snow pack.

Within the active layer a wide variation in ice content is present throughout an annual cycle of freezing and thawing /Kane and Stein, 1983/. Therefore it is a complicated task to predict the infiltration rates into frost active ground. Initial calculations of infiltration rates could be done with empirical formulas, such as Horton's- and Phillip's equations /Kane and Stein, 1983; Woo, 1986; Wilson, 1990/. /Kane and Stein, 1983/ concluded that the infiltration rate is inversely proportional to the moisture content, both liquid and frozen, in the ground.

The infiltration of meltwaters into the ground and subsequently the refreezing of the infiltrated water convect heat into the ground, which rapidly increases the ground temperature /Woo, 1986/. The infiltration into a frozen unsaturated ground can continue for a long time, but more frequently refreezing of the infiltrated water seals the pores and the infiltration temporarily ceases. If the meltwater is inhibited from infiltration into the frozen ground, it will freeze /March and Woo, 1984/ at the ground surface and while freezing inject a significant amount of heat into the ground.

4.1.3 Lateral groundwater flow

In the active layer, depression of the frost table during the thaw season depends on local thermal properties and microclimatic influences. Therefore the frost table will have an irregular surface causing a complex groundwater regime, with "pots and hummocks". These features create a groundwater system where not necessarily the ground surface topography governs the flow pattern /Woo and Steer, 1983/. However, the driving force in the active layer is as in traditional hydrology the hydraulic gradient. Therefore the available amount of water and the topographical nature govern the total flow through the system.

The annual water budget within the active layer is very varied. All hydrogeological conditions, from a tight formation, an unconfined to a confined formation, are found throughout the annual cycle. Further, the available amount of water is normally limited during the summer, and at times during the summer the active layer might even be drained of water.

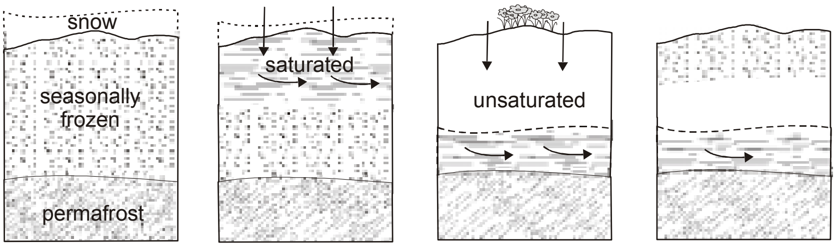

Figure 4-2. *Example of the annual water budget in an active layer from winter to autumn /from Haldorsen et al, 1997/.*

Much of the continuous permafrost regions (e.g. the tundra) is a tree-less terrain with a continuous organic (peat) layer. The soil profile in these regions is, in general, composed of a highly permeable layer of organic soil (peat) overlying mineral sediments with a significant lower permeability /Quinton et al, 2000/. If these mineral sediments are frost-susceptible, cryoturbation (cf section 5) may produce hummocks of mineral sediment (MacKay in /Quinton et al, 2000/). This patterned ground creates a network of high permeable pathways of peat with less permeable 'isles' of mineral sediments blocking an isotropic lateral groundwater flow /Quinton and March, 1998; Quinton et al, 2000/. The inter-hummock channels have a thickness of a typical order of 0.3 to 0.5 metre, and show a decreasing hydraulic conductivity with depth due to a higher decomposition of peat at depth /Quinton and March, 1998/. These factors together with discontinuities (e.g. pipes) present between the organic layer and the mineral sediment /Gibson et al, 1993; Carey and Woo, 1998, 2000/ are the main causes for the great variation in the lateral groundwater flow within the permeable bed that forms the suprapermafrost aquifer.

Slope hydrology

Slope hydrology is one of the most investigated parts of the Arctic hydrology /e.g. Woo and Steer, 1983; Hinzman et al, 1993; Woo and Young, 1993; Quinton et al, 2000/. Much of the annual run-off is one part of the slope hydrology with much drainage done in early spring, when the subsurface still is relatively tight and much flow is surface run-off /e.g. Roulet and Woo, 1988/.

Later in the spring and during the summer much, if not all, of the precipitation is able to infiltrate into the slope material and be discharged as subsurface flow in the active layer. /McNamara et al, 1998/ demonstrated that an Arctic basin responds quickly to precipitation events with extreme and intense run-off peaks. However, they also posed the problem with an extended recession of the run-off. This extended recession of the run-off is partly explained by the extensive storage capacity of the active layer but fails to give a satisfactory explanation why this storage is "by-passed" at the immediate start of the precipitation event. The only proposed and plausible explanation is that throughout the year large parts of the subsurface is so saturated that a surface run-off is triggered even at a small precipitation event.

/Åkerman and Malmström, 1986/ presented a theory on the creation of permafrost mounds as a result of subsurface slope hydrology (see Figure 4-3). This theory explains why these 'pingo-like' ice features are created at unfavourable locations such as at the end of a slope. These features can cause extreme pressures on structures located in the vicinity of the mound.

The theory presented by /Åkerman and Malmström, 1986/ is based on groundwater flow in the porous slopes. This flow follows Darcy's law. The ground water is normally of meteoric origin and may be stored as ground ice for long periods of no-flow conditions /Woo, 1986/. However, during the summers the active layer may transport large amounts of water. Later when the autumn comes the ground starts to refreeze from the top. This creates a confined layer where the groundwater pressure may increase. Trapped in the discharge area of the slope this groundwater may break through the confining layer and create icings (cf section 4.1.4) or develop into a mound containing an ice core.

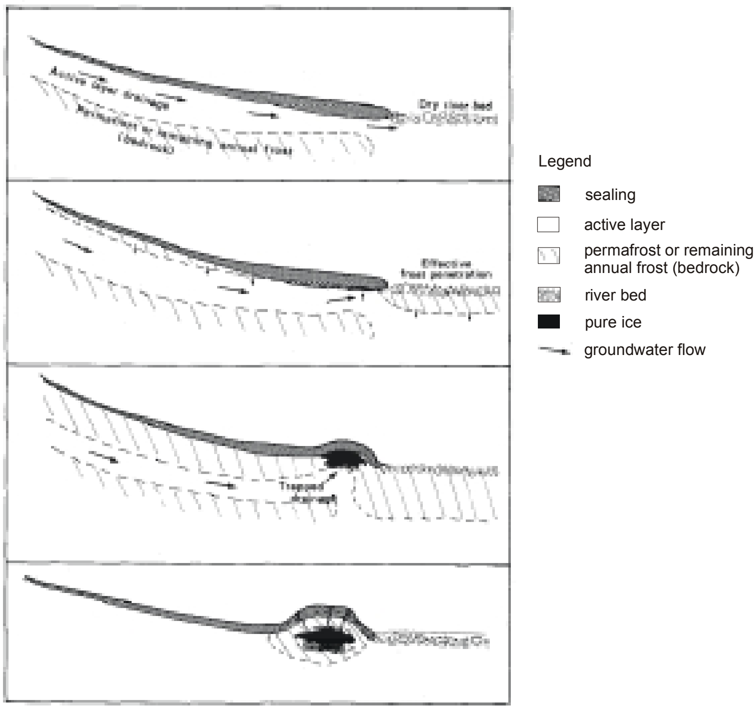

Figure 4-3. *Schematic sequence of events in the development of a permafrost mound /from Åkerman and Malmström, 1986/.*

Frost blisters are of similar origin as the permafrost mounds. The blisters can as well cause a significant pressure on its surroundings especially if fed from a perennial source of water. That may for example, be a hydrothermally open talik in a fracture zone.

In steep slope hydrology such as flood banks special criteria may impact. /Seppälä, 1997b/ suggested that water from heavy precipitation or melt events could cause piping between potential high-pressure water sources, where water is stored in the fringes of ice-wedges. He found much evidence of this kind of behaviour at flood banks in Canada.

4.1.4 Seepage

Groundwater seepage (discharge) from the active layer occurs where the water table rises above the ground surface. There are several conditions favourable for seepage; according to /Woo and Xia, 1995/ the following parameters control the seepage from the active layer: grain size, frost table, water supply, and topographical relief.

If the water source for the discharge is truly an active layer aquifer the discharge will fluctuate annually and may even perish during the cold period. But, if the seepage water is derived from a subpermafrost source or sometimes even deep suprapermafrost sources such as troughs of water (closed taliks) the discharge rate may even be constant throughout the year /van Everdingen, 1990; Haldorsen and Heim, 1999/

Icings
Surficial discharged water, in a cold environment, may freeze into sheet-like layers of ice on the ground. These layers of frozen ice create an icing. Essentially all icings are more or less related to a discharge of groundwater /van Everdingen, 1990/. If the discharge source is a suprapermafrost aquifer the growth of the icing will eventually perish, but if fed from deeper aquifers the icing will probably continue to grow all through the cold period /Church, 1974; van Everdingen, 1990/.

4.2 Hydraulic parameters of permafrost

4.2.1 Ground water content

The total ground water content depends on where in the subsurface the water exists, e.g. depth in relation to the surface but also the depth in relationship with the groundwater table has a pronounced impact on the ground water content.

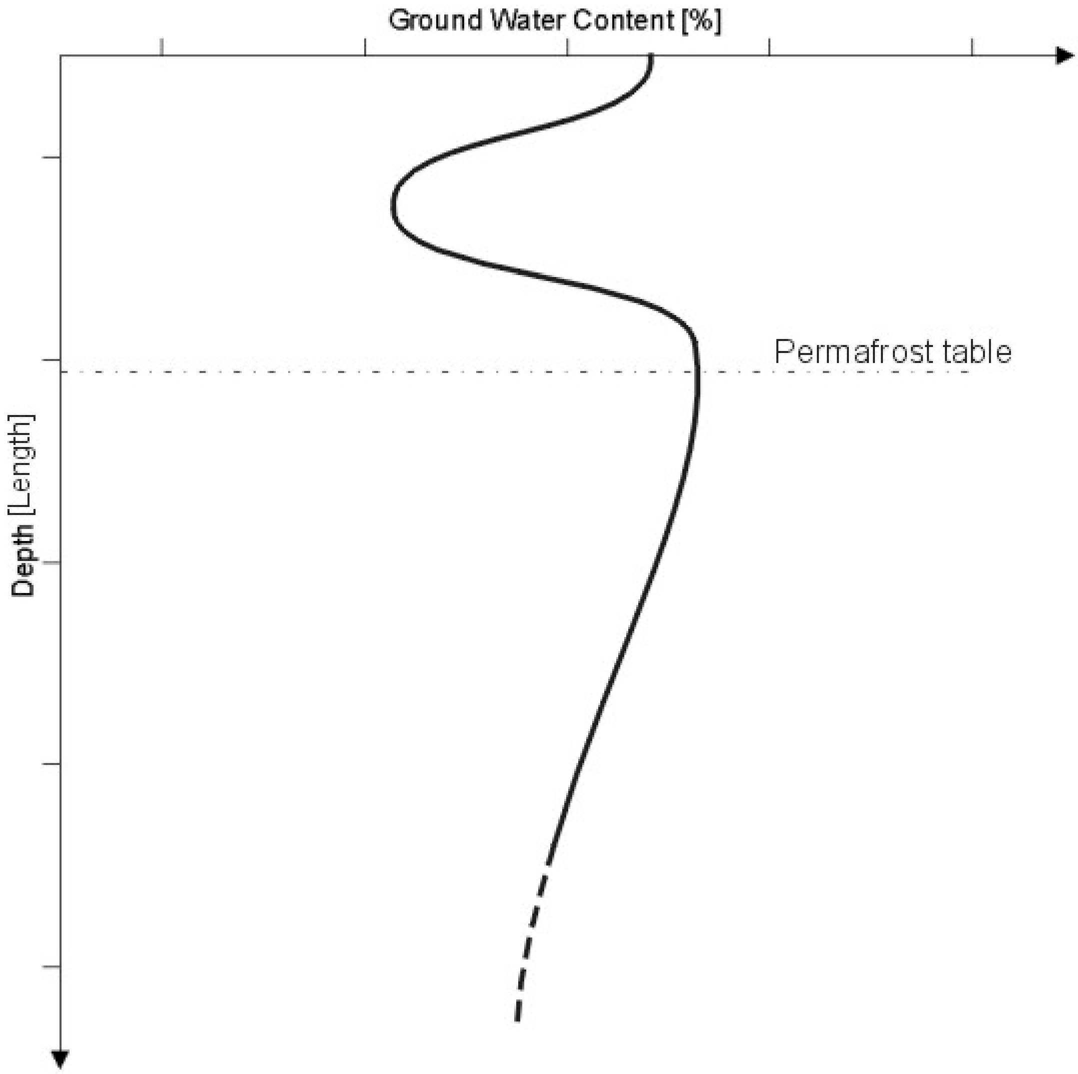

Figure 4-4. *Principal illustration of ground water content versus depth in an unsaturated frozen ground /based on Harris, 1988/.*

If the permafrost table is above the groundwater table, the processes of recharge and cryosuction will with time cause the groundwater table to reach equilibrium with the permafrost table. Cryosuction is the effect of suction created due to the development of a negative groundwater pressure, similar to the concept of capillarity. This negative groundwater pressure is due to the decreased amount of liquid ground water present in the frozen ground.

Water freezes at 0°C, that is a trivial statement, but in order to do so water needs to be pure, under a pressure of one atmosphere, and in sufficient amount /Williams and Smith, 1989/. That is, in most situations even a frozen ground constitutes of a certain amount of liquid water. Liquid water is present in the frozen ground as thin water films adsorbed onto the surfaces of ground particles. The thickness of this water film, hence the liquid ground water content, is a function of the temperature /Anderson and Morgenstern, 1973/ and is almost independent on the total ground water content (moisture, liquid and solid). The water film thickness is approximately 50 Å or more at 0°C, 9 Å at –5°C, and 3 Å at –193°C.

The quantity of liquid water present in frozen soil is further related to soil type. A fine-grained soil with large surface area has greater amount of liquid water than do a coarse-grained soil /Kane and Stein, 1983/. Further the ground water content has impact on both heat and mass transport in the ground /Romanovsky and Osterkamp, 2000/.

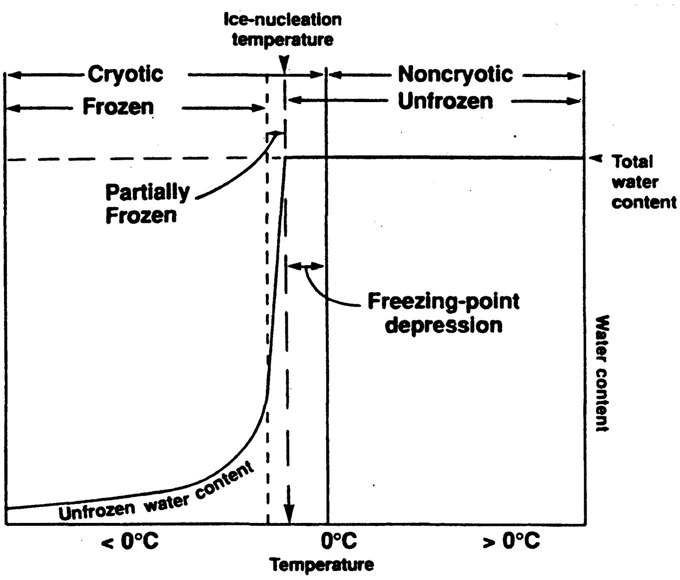

Figure 4-5. *Unfrozen ground water content and temperature /from van Everdingen, 1990/.*

4.2.2 Hydraulic conductivity

Even in a frozen ground the liquid water adsorbed to the surfaces of the ground particles is mobile under the influence of a gradient. In all groundwater systems the movement is from a position of high energy towards areas with low energy. The distinctive driving force of water movement in saturated frozen ground is the thermal gradient /Harlan, 1974; Williams and Smith, 1989/ and, in general, gravitational and hydrostatic forces are negligible. In such a system the liquid waters' mobility is described with the regelation theory, that describes a mass transfer process governed by a thawing and freezing front moving along with the latent heat flow /Kane and Stein, 1983/. However, alternative views are present; /Burt and Williams, 1976/ measured the hydraulic conductivity of frozen ground by the use of a simple permeameter. Using Darcy's law and a controlled pressure difference they concluded that frozen ground is permeable but also that the hydraulic conductivity was significantly reduced with decreasing temperature and that the presence of ice lenses further reduced the hydraulic conductivity (cf section 4.2.1 third paragraph).

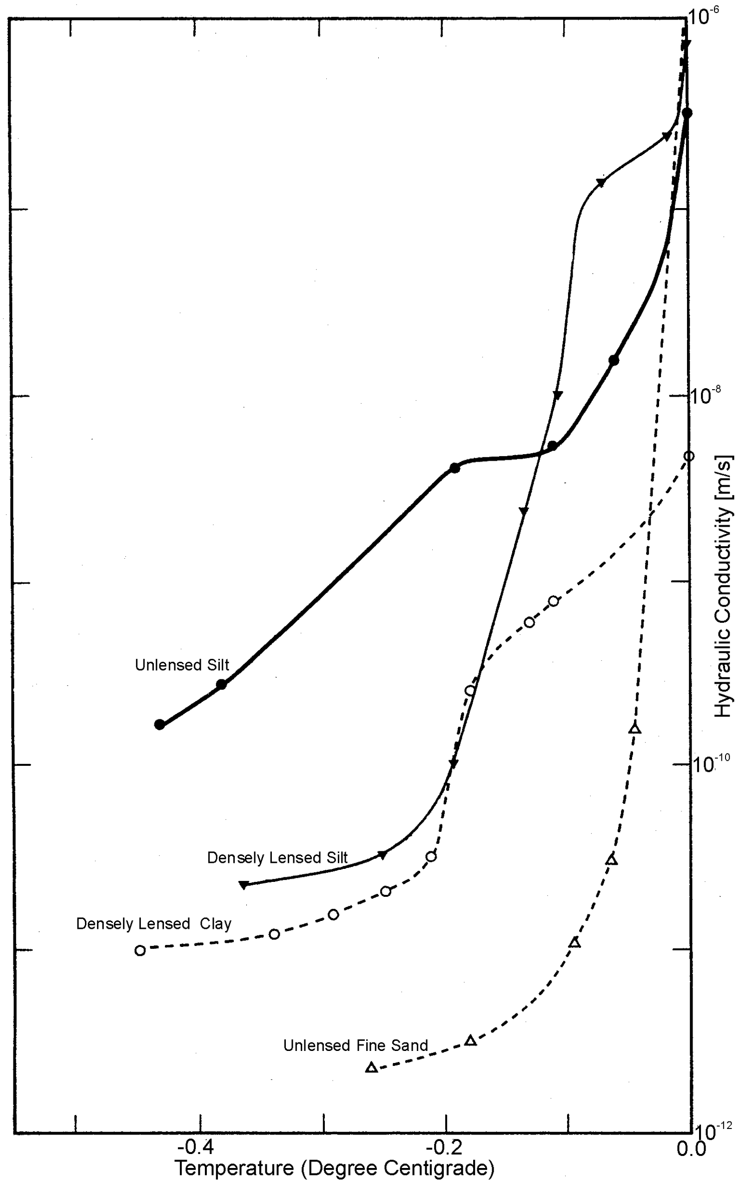

***Figure 4-6.** The relationship between temperature and hydraulic conductivity for different saturated frozen soil types /from Burt and Williams, 1976/.*

At a workshop on “the impact of climate change & glaciations on rock stresses, groundwater flow and hydrochemistry” /King-Clayton et al, 1997/ it was concluded that due to the paucity of hydraulic conductivity data there was no consensus on the hydraulic conductivity of permafrost ground. Presenting teams were found to use hydraulic conductivity values of approximately ten per cent of the initial ground value, values corresponding to clayey silts, and some assumed the permafrost to be impervious. Although debated there was a general opinion that the hydraulic conductivity should not be considered stationary and equal to zero, but time-dependant and of significance to fluid flow /King-Clayton et al, 1997/.

4.3 Groundwater flow in permafrost regions

The groundwater flow is substantially restricted by permafrost and may only take place within the active zone or at greater depths within open taliks or beneath the permafrost. The active layer is mainly responsible for local flow within slopes (cf section 4.1). Therefore the large-scale groundwater flow is restricted to deep aquifers. The presence of a permafrost layer will force the lateral flows toward greater depths /McEwen and de Marsily, 1991/. The limiting effect on recharge caused by the ‘impervious’ permafrost layer together with a deeper flow regime will cause a decrease of the groundwater flow that otherwise would flow through the deep unfrozen parts of the bedrock /McEwen and de Marsily, 1991/. Importantly, even though the amount of deep groundwater recharge in the continuous permafrost regions is limited, the most active water exchange and groundwater fluxes in these regions occur in subpermafrost aquifers /Yershov, 1998/.

Recharge to deep groundwaters will be significantly affected by the occurrence of permafrost. Extensive peat bogs, numerous lakes and ponds in regions with low rainfall illustrate this restriction of recharge especially in areas of low topographical variation /Roulet and Woo, 1988; McEwen and de Marsily, 1991/. In present day Spitsbergen (part of the Svalbard archipelago), glaciers cover much of the land surface. In unglaciated areas the extensive permafrost prevents a recharge of deep groundwaters. Therefore, as in many Arctic environments, this recharge is restricted to be sub-glacial /e.g. Haldorsen and Heim, 1999/.

The only possible locations for a deep groundwater recharge in permafrost areas are at conductive passages through the permafrost, such as taliks. Taliks of variable distribution and connectivity underlie many of the larger lakes and rivers. If such taliks form a continuous unfrozen zone between the active layer and the subpermafrost aquifer, the surface waters may develop into a zone of either recharge or discharge of groundwater /e.g. Kane and Slaughter, 1973; McEwen and de Marsily, 1991/. In the Huolahe Basin (Northeast China) the subpermafrost groundwater exists throughout the entire groundwater basin and taliks are the main path for recharge of the subpermafrost aquifer. Here, all taliks are located in structurally deformed zones in the bedrock and form both recharge zones as well as discharge zones /Fengton and Guangzhong, 1988/. The almost impermeable permafrost layer on top of the Huolahe Basin varies between 70 and 130 metres. According to Fengton and Guangzhong, the Huolahe Basin creates, through the permafrost’s confining behaviour, an artesian groundwater system with over-pressures as high as 10 metres of water with groundwater pressures varying linearly from the groundwater divide down towards the discharge areas.

Many researchers have concluded that during the Weichselian glaciation the proglacial regions were cold and dry /e.g. Velichko, 1984; Dawson, 1992; European Commission, 1997/. Under such conditions it can be assumed that the sources for water are sub-glacial melt, permafrost melt, and during thawing input of superficial snow and ice melt /Boulton and Curle, 1997; van Weert et al, 1997/. On the basis of results from two-dimensional groundwater models it was concluded that the entire proglacial zone is over-pressurised and all present discontinuities in the permafrost functions as groundwater discharge areas. The discharge of the subpermafrost aquifers on Spitsbergen occurs in the ocean or in well-defined taliks /Haldorsen and Heim, 1999/ which are all found below the highest level for the postglacial shoreline. In Canada active groundwater systems in permafrost regions have been deduced from discharge phenomena. As in non-frozen environments these phenomena include springs, baseflow to streams, and hydrochemical annual variations /van Everdingen, 1974/. In 1988 /van Everdingen, 1988/ reported that a spring area in Yukon had an average discharge rate of approximately 360 litres per second and was composed of snow meltwater and rainfall waters that infiltrated the system some 15 to 20 years earlier. The permafrost in this area is relatively thin with depths of approximately 5 to 10 metres.

An implication of the results from numerical modelling of the proglacial regions for the Weichselian period could be that the deep groundwater recharge of meteoric waters is inhibited by the proximity of an ice sheet.

Pingos may look as small volcanoes and can be 50 metres in height. Pingos are a permafrost feature that is believed to form at the site of a talik. The pingo is, in general, constantly fed with pressurised water from a talik. From a genetic view the creation of pingos depends on a pressurised groundwater system, with a constant recharge of water to a frozen core of ice. Pingos could, as traces of a talik, be one descriptive part of the regional groundwater flow.

The presently known taliks have rather constant extensions and are not subject to seasonal freezing and thawing /van Everdingen, 1990/. Further new taliks are constantly created due to different human activities /Yershov, 1998/.

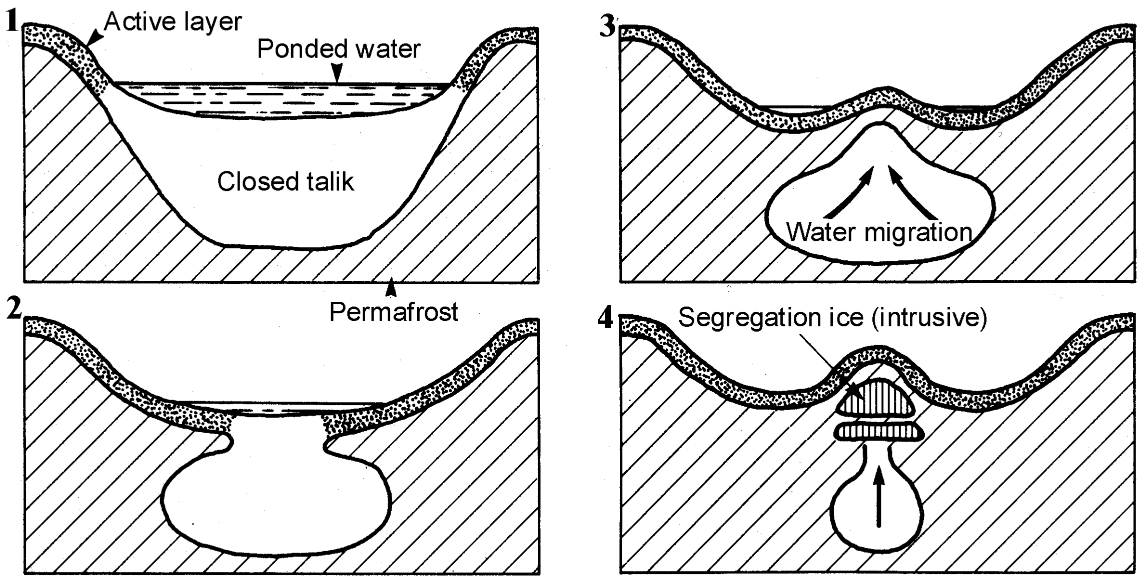

Figure 4-7. *Stages in development of pingos /from Williams and Smith, 1989/.*

5 Mechanical effects in permafrost regions

The temperature in the ground will affect the mechanical strength of the geological subsurface. From the observed fact that the compressive strength of a wet material increases more than that of the same but dry material, it can be concluded that the water content in pores and fractures has a significant control on the strength of the material /Mellor, 1973; Kuriyagawa et al, 1980; Dahlström, 1992/. One of the pronounced physical changes associated with the freezing of ground waters is the volume expansion /e.g. Williams and Smith, 1989; Ahonen, 2001/. The specific volume of ice is approximately nine per cent larger than that of water. Further when water is transformed into ice the material experience an additional strength due to the enhanced cohesion caused by the internal strength of ice.

The strength of permafrost ground decreases with the duration of an applied load and with the exposure to the atmosphere /Vakili, 1993/. This is due to loss of the additional cohesion and the characteristic creep-behaviour of ice. Applied loads on permafrost ground cause creep in the elastoplastic ice. Thawing of ice decreases the additional cohesion, which will subsequently disappear when the ice is fully transformed into liquid water. This loss of strength will in soils sometime lead to slope stability problems /e.g. Williams and Smith, 1989; French, 1996/.

Cryoturbation is the collective term for all movements in ground caused by a frost action. Cryoturbation encompasses e.g. frost heave and thaw settlement as well as expansion and contraction due to temperature changes, irrespective of perennial or seasonal change. The presence of water and an additional phase change is necessary in order for a cryoturbation to arise. Cryoturbation is an important process in the development of patterned ground.

Solifluctions is the used notion for slow down slope flow of saturated unfrozen earth materials. Solifluction is not restricted to areas where frost action prevail. Further solifluction has no dependence on a frozen substrate. The active layer is however, a potential risk layer for solifluction processes, since the active layer during certain times are well saturated. Furthermore, creep of frozen ground is one of the solifluction processes.

The rock mechanical properties of coal and shale were investigated in a mine on Spitsbergen /Myrvang, 1988/ and it was concluded that there were little or no difference between the frozen and unfrozen properties. Myrvang, did however, observe that the roof of the mine deteriorated. However in areas where shotcrete was applied no such deterioration could be observed. Hence, the conclusion drawn was that the deterioration depended on the contact with moisture and air and not primarily on the thawing. On the extreme end of the freezing process (that is very cold environment, and very large temperature gradient), experiences from refrigeration of bedrocks for gaseous storage have shown that the in-situ stresses in the frozen rock mass surrounding the cavities have some tensile components. This stress situation is a result of the thermal contraction of the rock minerals. Tensile conditions within the rock mass decreases the normal stresses acting on fractures and hence enhance the risk of block instabilities.

Frost heave and subsidence

Whenever the ground water contained in a soil freezes, the soil surface heaves and later when the soil thaws again the soil becomes soft and muddy /Førland et al, 1988/ mainly due to the over-saturation of the soil. These phenomena are called frost heave and subsequently frost subsidence and cause serious problems in construction engineering. As a rule of thumb a porous material with a void diameter of approximately 50 μm is the most frost susceptible material /Van Vliet-Lanoê, 1998/. This is because the curvature of the pores in such a material is large enough to yield the necessary lowering of energy levels that is significant for formation of segregation ice.

The segregation of ice is the result of cryosuction separating uniformly distributed ground water into ice lenses. In order for the formed ice lenses to grow in the ground the voids need to be continuous for movement of liquid water /Williams and Smith, 1989; Van Vliet-Lanoê, 1998/. This process is restricted to surficial layers due to a pressure dependence and therefore one seldom finds ice lenses at depths larger than 40 metres /Washburn, 1979; Van Vliet-Lanoê, 1998/. The frost heave is a result of the segregation process in addition to the volume expansion of water while freezing to ice. In bedrock the concept of frost heave is used but does in general describe one part of the frost wedging process (cf section 5.1.1).

Patterned ground

Patterned ground is a comprehensive term for different patterns in nature that occur due to the presence of permafrost. Patterned ground is often spectacular and examples are earth circles, stone polygons and rings, and soil- and ice-wedges /e.g. Williams and Smith, 1989/. Of these, polygon or wedge features are most interesting from a mechanical point-of-view. These latter features arise if the frost or thermal contraction fractures that may develop in frozen ground later are filled up with water (ice) and/or debris of different origin and size.

Thermal contraction fracturing occurs in most materials. A saturated soil expands while freezing as long as there is available water. However, after some time and below a certain temperature the freezing of the water is completed and the thermal contraction of the soil dominates the deformation /Williams and Smith, 1989/. The process of thermal contraction is complicated by the apparent creep deformations in soils and ice. The creep reduces the

Figure 5-1. *Fully developed stone circles (each circle is approximately 1–2 metres in diameter) (reproduction of a photo by Bernard Hallet taken from /Williams and Smith, 1989/).*

stresses in the soil and hence counteracts the thermal contraction that also tries to reduce the stresses. Therefore, in general, the soil needs to be exposed to the freezing process rapidly enough for creeping not to occur. Simultaneously the temperature needs to drop towards at least –6°C /Williams and Smith, 1989/ in order for the ground to contract.

Frost (contraction) fractures rise in a systematic manner and creates vertical fracture polygon networks in the ground. The size of the polygons, or circles as they may appear as, depends partly on the grain size of the soil.

Fine-grained soils cause a small 'radius' polygon and vice versa. If these thermal contraction fractures create small depressions in the ground surface they may with time accumulate the larger grains from the soil, moved by the 'frost pull' process (uplift of stones, see e.g. /French, 1996/), towards the surface and the topographical depressions. This stone accumulation creates the spectacular features of stone circles.

If the frost cracks instead get filled with soil or water they may with time develop into soil- or ice-wedges. Most favourable conditions for ice-wedges are to be found in lowlands, with a poorly developed drainage, within the continuous permafrost zone. However, ice-wedge networks are also found on well-drained outwash plains /Gray and Seppälä, 1991/.

Larger mounds in the permafrost region could be a pingo (cf section 4.3) or a palsa. Palsas are the result of cryosuction and the development of segregation ice, normally under a layer of surficial peat. Although the genetic definitions of palsas and pingos are firm /Worsley et al, 1995/ some permafrost features that seem to be an intermixture of these occur /e.g. Åkerman and Malmström, 1986/. The transitional geomorphologic forms (permafrost mounds) described by /Åkerman and Malmström, 1986/ (cf section 4.1.3) are all occurring at the down-slope end of slopes where seepage of lateral groundwater flow within the active layer is inhibited by the presence of an annual frost table. Traditionally, pingos were considered to be features of continuous permafrost areas, while palsas were believed to occur in areas of sporadic or possibly in discontinuous permafrost regions /Worsley et al, 1995/. However, reports on palsas in continuous permafrost /e.g. Åkerman, 1982/ have been presented.

5.1 Weathering and Erosion

The weathering characteristics in tropical environments are relatively well understood, with its penetration of waters into the ground followed by the hydrolysis and the breakdown of feldspar minerals into clay minerals. In cold climates weathering processes are not as well known. In general, the major active processes are believed to be frost wedging and wind erosion and possibly exfoliation due to thermal loads /French and Guglielmin, 2000/.

5.1.1 Frost wedging

Frost wedging is the result of the physico-chemical process where water expands by approximate nine per cent of volume while freezing to ice. As doing so, the water exposes the surrounding for an extreme pressure that may rupture the ground matter, but more often expand existing fractures and/or porous media porosity. Frost wedging can also cause impact such as rock displacement.

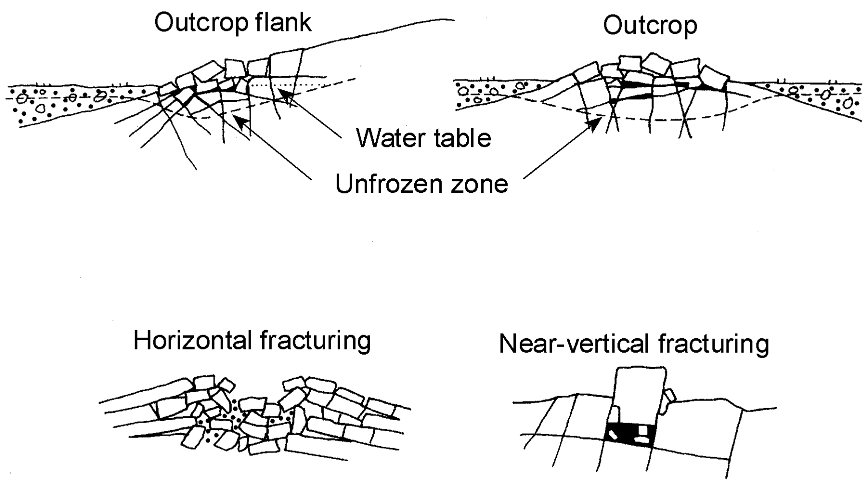

Figure 5-2. *Examples of "bedrock heave" (frost wedging) /from French, 1996/.*

Frost wedging occurs on all scales. In the micro-scale, frost wedging plays an important role as a main part in the cryotic weathering.

5.1.2 Wind erosion

Compared to moving ice or running water, the wind is a relatively insignificant erosional agent. However, if abrasive materials are available the wind will cause significant erosion. In permafrost regions with extensive vegetation, the vegetation effectively binds the abrasive materials. Therefore only the bare regions of polar deserts are exposed to a significant amount of wind erosion /French, 1996/. The wind is also responsible for the transportation of aggressive species, such as salt, which may enhance e.g. weathering.

The process of deflation will with time create a 'pavement' that will protect the permafrost land from further erosion. The deflation takes away all fines and leaves a 'pavement' of larger grain-sizes that sits by gravity. With time the ceasing of the deflation yields less and less abrasive material and the subsurface is further protected from erosion.

6 Hydrogeochemistry in permafrost regions

Freezing within the ground will change the hydrogeochemical equilibrium in comparison with unfrozen conditions. At lower temperatures some substances have less solubility and others will have increased solubility. The same is true regarding dissolution of minerals. A permafrost layer will partly inhibit an infiltration of meteoric water or possibly glacial meltwaters. This 'impervious' layer will therefore change the flow system, the fluxes and hence the hydraulic and chemical processes active within the ground.

In the work performed by the nuclear waste management companies the origin of observed saline groundwater has been considered in detail /e.g. Laaksoharju et al, 1999/. Some researchers believe the salinity of the groundwater may, at least partly, result from salt rejection due to freezing /Gascoyne, 2000/.

Diluted unfrozen waters in permafrost regions may be found if the water is derived from /Gascoyne, 2000/:

- the active layer where the residence time is short,
- degrading permafrost,
- flow systems with high flow velocities (e.g. karstic, fractured).

In permafrost regions in coastal and marine areas, and in deep environments, however, the composition of unfrozen water is usually more saline.

The groundwater chemistry depends on the hydrogeology of the aquifer and the source of the water and therefore has large variations. But, in general, subpermafrost waters have similar composition as that of overlying suprapermafrost aquifers and taliks /McEwen and de Marsily, 1991/.

A comprehensive review of hydrogeochemical effects of freezing, that is permafrost, is found in /Gascoyne, 2000/ and /Ahonen, 2001/.

6.1 Weathering

The chemical weathering in cold environments has received relatively little attention. It is, however, one part of a more complex physico-chemical process /French and Guglielmin, 2000/. In this process the mechanical effects, and especially the frost wedging at different scales, are believed more important in crystalline hard rocks (cf section 5.1).

The dissolution of carbonate minerals and the silicate weathering are temperature dependant reactions. Low temperatures slow down the chemical reactions, but they do not cease /Appelo and Postma, 1996/. Based on research at Spitsbergen (Svalbard) /Åkerman, 1983/ has found that at present the local limestone denudation is a couple of millimetres per one thousand years.

6.2 Salt rejection

The ice crystal is not able to accommodate dissolved chemical species. As a consequence the liquid phase will experience an aggradation of chemical components, of which salt has been the most thoroughly investigated. This aggradation of salt in the unfrozen groundwater is commonly referred to as salt rejection. However, other species than salt, now bounded in e.g. fracture fillings, could possibly be used as proxies of historical permafrost distributions /e.g. Gascoyne, 2000/.

One consequence of the salt rejection process is that the freezing front of permafrost is 'pushing' a supercooled, saline liquid in front of itself /e.g. Ahonen, 2001/. Such a pushed saline liquid will have an impact on the freezing point more pronounced than that of increased pressure conditions (cf Figure 6-1). As quoted by /Ahonen, 2001/: " If assumed that the original salinity of groundwater in the bedrock is about 10 g/l, which is a typical value at a depth of about 400–500 metres. If 90 per cent of the water freezes slowly allowing the residual fluid to accommodate the dissolved salts, salinity of the residual fluid will be about 100 g/l. It can be estimated that the freezing temperature of such solution is about –5°C."

This kind of salt rejection creates a basal cryopeg that may be part of the brines located at depth, but may as well float on top of a fresh water aquifer. In this latter case the saline-/fresh water interface is in an unstable mode since the heavier fluid sits on top of the lighter fresh water.

An increasing salinity in the unfrozen ground water could further lead to a "supersaturation" regarding the mineral phases. Vogt /in Ahonen, 2001/ reported on minerals such as, calcite, silicates, and (Fe, Mn) oxides, that have been found precipitated in cryogenic environments.

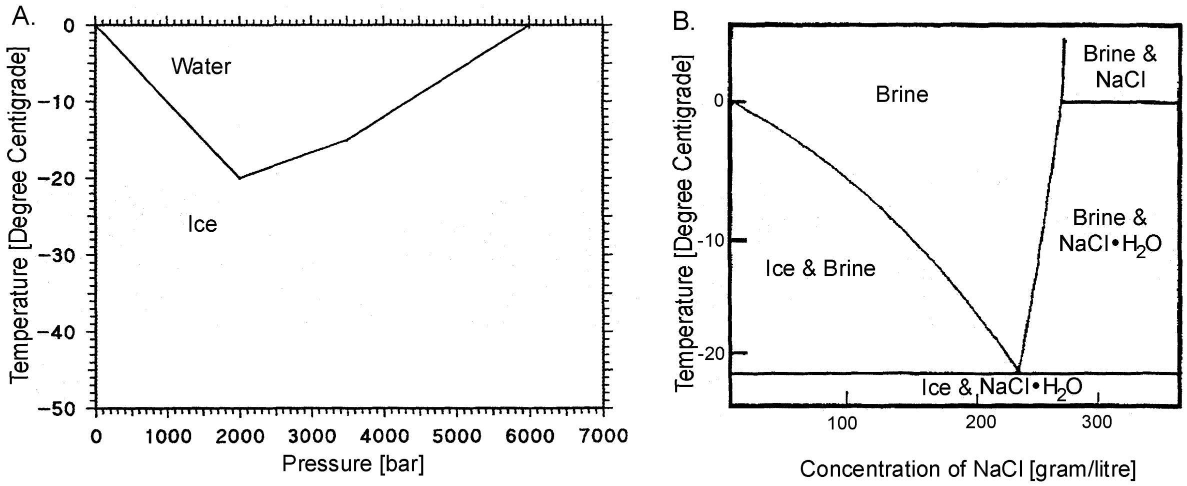

Figure 6-1. *Part A. Illustration of the pressure dependence on the phase diagram for the ice/water system. Part B. Illustration of phase diagram for the NaCl/H_2O system /from Ahonen, 2001/.*

6.3 Gas hydrates

Gas hydrates – also referred to as clathrates – are gas molecules of different kinds enclosed in the ice crystal structure. Methane hydrates are widespread in sub-sea permafrost but also common in continental areas. The source of methane is either biogenic or thermogenic /Ahonen, 2001/. Of biogenic origin is the typical hydrates found at the seabed. These hydrates are formed slowly and over waste areas as bacteria digest organic matter that have sunk to the bottom of the sea. Thermogenic hydrates are formed from escaping gases from the interior of the Earth. These hydrates are believed to be more concentrated and localised.

Gas hydrates occur, in general, in ground below a depth of approximately 200 metres and can theoretically be expected down to depths of 1.8 kilometres (Judge in /Gascoyne, 2000/). In the Mackenzie Delta, Canada, gas hydrates typically are found in coarse grained sandy intervals, with the fine grained interbeds being non-gas hydrate bearing /Dallimore and Collett, 1998/. The gas hydrate content and stability are dependent on the gas chemistry, the pore pressures and the salinity of the groundwater.

The growth of gas hydrates in the ground excludes salts /Gascoyne, 2000/ and also results in an isotropic enrichment in the solid phase of approximately the same amount as in the ice-liquid water fractionation /Clark et al, 1999/. Gas hydrates form in high pressure/low temperature environments when water and gas exist together. Such conditions are typically found in permafrost regions where methane hydrates also are found abundantly /Ahonen, 2001/.

Figure 6-2 illustrates a phase diagram for pure methane hydrate/methane gas. As the figure indicates, the stability of the hydrate is pressure dependent. The minimum pressure for methane hydrate to form at 0°C is at a hydrostatic pressure of about 200 metres of water. If instead the temperature goes up to 15°C the necessary depth is approximately one kilometre.

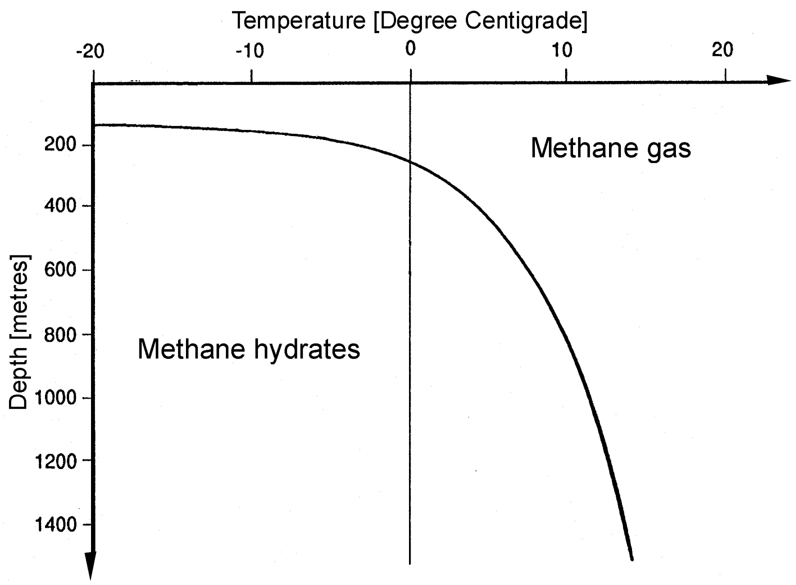

Figure 6-2. *Illustration of the depth/temperature relationship for Methane hydrate /from McEwen and de Marsily, 1991/.*

A permafrost layer may seal off escape routes for gases within the subsurface. Hydrates in Swedish bedrock could only have existed trapped in the porosity provided by fractures /Talbot, 2000/. The trapped gases are prevented from rising higher but could also be bounded in solid clathrate structures, since temperatures are lower in the permafrost region and hydraulic pressures may be affected by upstream recharge. Such trapped hydrates have the potential to volatilise explosively if exposed with a sudden pressure drop /Talbot, 2000/.

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Appendix I

Some departments, institutions, or societies that study permafrost and associated sciences

Arctic Climate System Study (ACSYS) project
(http://acsys.npolar.no)

Arctic Methane Hydrate Research Well Program
(http://sts.gsc.nrcan.gc.ca/page1/hydrat/content.html)

Circumpolar Active Layer Monitoring program
(http://www.geography.uc.edu/~kenhinke/CALM)

Earth Cryosphere Institute, Russia
(http://www.ikz.tmn.ru/Ikzeng.html)

Geological Survey of Canada
(http://sts.gsc.nrcan.gc.ca/permafrost/)

Institute of Low Temperature Science, Hokkaido University
(http://www.lowtem.hokudai.ac.jp/english)

Institutionen för naturgeografi och kvartärgeologi, Stockholms Universitet
(http://www.natgeo.su.se/ink/home.html)

International Permafrost Association
(http://www.geodata.soton.ac.uk/ipa)

International Union for Quaternary Research
(http://inqua.nlh.no/) and (http://inqua.nlh.no/permafrost.html)

Naturgeografiska institutionen, Lunds Universitet
(http://www.natgeo.lu.se/)

Permafrost in Alaska from Fairbanks, Alaska
(http://fairbanks-alaska.com/permafrost.htm)

School of Geography and Geology, McMaster University, Ontario, Canada
(http://www.science.mcmaster.ca/geo/geomain.html)

Snow, Ice, & Permafrost Group, Geophysical Institute, University of Alaska Fairbanks, USA
(http://www.gi.alaska.edu/snowice/)

State Hydrological Institute, St Petersburg, Russia

United States. Permafrost Association
(http://www.uspermafrost.org)

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