Transient thermal modeling of palsa distribution in Northern Norway, Finnmark

And the importance of estimating local factors controlling palsa development.

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Thesis submitted for the degree of Master of Science in Physical Geography 60 Credits

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Transient thermal modeling of palsa distribution in Northern Norway, Finnmark. And the importance of estimating local factors controlling palsa development.

A thesis submitted by Ingvild Solheim for the degree of Master in Physical Geography.
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Supervisor(s): Sebastian Westermann (UiO) - Bernd Etzelmüller (UiO) and Karianne Staalesen Lilleøren (UiO).

Front page: Presenting a degrading palsa in Karlebotn, Finnmark, and the surrounding vegetation affected by the thawing of the ground ice within the palsas. Notice the dead birch trees in the thermokarst lake. Bubbles of methane popped up when disturbing the water with a pole.

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Summary

Palsas and peat plateaus are permafrost mounds situated in mires with a core of ground ice. In northern Norway, palsas are located in the sporadic permafrost zone, and demarcate the southernmost limit of permafrost. Thus these landforms are vulnerable to climate change. Polar Regions contain immense amounts of organic carbon frozen in the soil. Permafrost degradation in these areas leads to emission of greenhouse gasses and may further accelerate the climate feedback on hydrology, vegetation, infrastructure and climate. Permafrost models have proven to be sufficient tools in predicting the future scenario on ground temperatures and climate change. However, such models implement parameterizations on soil, vegetation and climate data which encompass broad uncertainties, as land cover and soil stratigraphy are estimated on a coarse scale and interpolated for large regions. Moreover, permafrost is controlled by local factors, which may vary over small scales. Therefore, sufficient datasets and parameter initializations are crucial in order to reproduce the state of the art- and future ground temperature regimes in permafrost regions.

The grid based transient thermal model CryoGrid 2 has been forced with stratigraphy- and snow scenarios based on in situ measurements on both a local- and regional scale for palsa mires in Finnmark. The resulting ground temperatures and thaw depths are in surprisingly well agreement with in situ GSTs and active layer depth for the region, and validated against a recent map over the palsa distribution in Finnmark. The presented soil and snow scenarios may benefit permafrost models in improvement on parameters which are controlling the temperature regimes for permafrost in organic rich soil.

Moreover, palsa distribution in Finnmark is strongly dependent on the insulating organic peat layer, and low snow depth throughout the winter in order to survive the current climate. The average thaw depth for the widely distributed study sites are comparable with one another and measured to be roughly 0.5 m.

The presented work may be a small step towards an improved representation of the circumpolar ground climate.
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Without doubt, the last year has been the most valuable year as a student, and there are many people who deserve my greatest appreciations. A large part of this thesis contains results from modeling, which I most likely would neither have been able to, nor would I have imagined it to be so intriguing, without great guidance and motivation from my supervisors Sebastian Westermann and Bernd Etzelmüller. They had the main ideas behind this project and introduced me to permafrost modeling during an incredible course in Japan last year. I will never forget one of the first discussions around this master project I had with Bernd, who said something like; people need to stop thinking they’re not capable or good enough in doing things they have not done before, and defy the controversial “Janteloven”, believe it can be done- especially in science.

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Thank you Karianne, Amund, Mari, my big sister Silje and Mads for helping in the review process in the final stages of this project.

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Blindern, June 1.st 2016

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1 Introduction

1.1 Permafrost and Climate

As we are facing great challenges on climate change and geohazards, an accelerated knowledge and interest regarding the cryosphere, climate change and its consequences on the Earth is acquired. According to UN’s “Intergovernmental Panel on Climate Change” (IPCC, 2014), the globally averaged combined land and surface temperature have increased with 0.85 °C from 1988 to 2012, and it is unequivocal that the climate on the Earth is warming. The Polar Regions, Arctic and subarctic regions have been considered most vulnerable to climate change, as the largest temperature increase has been observed in higher latitudes (Hanssen-Bauer and Førland, 1998, Førland and Hanssen-Bauer, 2003, Hinzman et al., 2005, Overland et al., 2008). The reported warming of permafrost temperatures in the Nordic area has been discernible since the beginning of this millennium (Christiansen and Etzelmuller et al., 2010). Such trends are expected to continue in future decades (Lawrence and Slater et al., 2008a,b, Zhang and Wang et al., 2008, Etzelmüller and Schuler et al., 2011, Hipp and Etzelmüller et al., 2012, Etzelmüller, 2013, Farbot and Isaksen et al., 2013). As a consequence, permafrost temperatures as well as active layer thickness are expected to increase. This has been documented for several regions (Romanovsky and Osterkamp, 1995, Osterkamp, 2005, Lemke and Ren et al., 2007, Jiang and Zhuang et al., 2012). However, according to Lemke and Ren et al. p. 370 (2007), some areas have been experiencing less warming and cooling as well.

One of the key components in the cryosphere, permafrost, has generated substantial interest during the last decades and the awareness continues to evolve among today’s society, due to the linkage it has to our global climate system, infrastructure, and geohazards (Haeberli and Guodong et al., 1993). Permafrost, a thermal condition of the ground, is defined as soil which does not exceed a mean annual surface temperature (MAGT) of 0°C during at least to consecutive years (Muller, 1943, Harris, 1986, Risborough et al., 2008, IPCC, 2014). This temperature dependent phenomenon affects the global climate system with its influence on the carbon budget, energy exchange, geohazards and hydrological processes (Cao and Gregson et al., 1998, Marushchak and
Brown and Romanovsky (2008) states that permafrost is classified as one of six cryosphere indicators of climate change, and it is also incorporated in the Global Climate Observing System (GSOS) as an essential climate variable for terrestrial systems (GSOS, 2012). In 1990 IPCC stated that research on the connection between permafrost and climate, including the effects of temperature forcing from climate variation, local environmental factors such as snow, vegetation and lithological conditions should be actualized and implemented (IPCC, 1990).

Recently, several permafrost mapping campaigns have been commenced in Europe, as a framework for the European PACE project (Harris and Haeberli et al., 2001), and The Global Terrestrial Network for Permafrost (GTN-P) (Burgess and Smith et al., 2000) and during the “International Permafrost Year” (IPY) several permafrost projects were established, e.g. in Antarctic (ANTPAS – “Antarctic Permafrost and Soils”) (Vieira and Bockheim et al., 2010), Russia (Romanovsky and Smith et al., 2010) and other sites in the Northern hemisphere (Christiansen and Etzelmuller et al., 2010) and in Asia (Zhao and Wu et al., 2010, Wu and Zhang, 2008, Wu and Zhang et al., 2012). Such campaigns have resulted in numerous studies in permafrost related research, and endorsed the understanding of the state of permafrost in the present and future scenarios. However, key research questions regarding the vulnerability of permafrost to thawing, the projected decline in permafrost extent in the near future, and the consequences it will have on ecosystem feedback and hydrological and infrastructural aspects, not to mention the possibility on a global scale with climate change due to remobilization of carbon pools from thawing permafrost still requires answers (Grosse and Romanovsky et al., 2011).

Permafrost occurs in different types of topography and forms and consists of soil, gravel and sand usually bound together by ice. And as permafrost is a temperature dependent phenomenon, and form in different types of soil and in different landscapes, and even subsea, it is common in very different landforms and scales, but not necessarily visible on the ground surface. However common forms which may be visibly interpreted as permafrost (active and relict) are numerous, e.g. rock glaciers, patterned ground, ice wedges, solifluction lobes, palsas, pingos and thermokarst lakes. Lowland permafrost regions are divided into continuous, discontinuous, sporadic, isolated and relict permafrost zones defined by the areal extent of permafrost within the landscape (IPA,
A total of 24% of the Northern hemisphere land surface is underlain by permafrost (Brown and Ferrians et al., 1997), and as mentioned, the future extent of permafrost is predicted to degrade, this will incur significant social and economic costs (Brown, 2001), not to mention the alarming effect of accelerated emissions of carbon dioxide and methane from permafrost in organic rich soils (Zuidhoff Kolstrup, 2000, Christensen and Johansson et al., 2004, Frey Smith, 2005, Grosse and Romanovsky et al., 2011, Kirpotin and Polishchuk et al., 2011, Pokrovsky and Shirokova et al., 2011, Schuur and McGuire et al., 2015).

Permafrost in organic rich soils occurs in vast areas of the Northern hemisphere, and landforms indicative of this are palsas and peat plateaus. These landforms are the most common landforms indicating permafrost in Fennoscandia with a widespread abundance mires in Norway, Sweden and Finland (Seppälä, 1986); however it is most abundant in the Western Siberian tundra where over 434,000 km² is covered in such peatlands. These palsa bogs are consisting of cores of excess ground ice overlain by rich organic sediments which constitutes perennially frozen peat plateaus or palsas. The height of these permafrost landform vary from one to several meters and the lateral extent may be up to several hundred square meters, whereof the larger palsas form peatlands (Kirpotin and Polishchuk et al., 2011). An increase in carbon fluxes to the atmosphere from the thawing of the organic carbon within these organic rich permafrost landforms could turn the subarctic into a net carbon source (Koven and Ringeval et al., 2011, Schuur and McGuire et al., 2015). However, ground temperatures in Fennoscandia are warmer than in Russia; hence the observed changes in palsa degradation in these areas and Northern America (Zuidhoff Kolstrup, 2000, Krüger and Leifeld et al., 2014, Borge and Westermann et al., 2016) may be an indicator of the future development of the immense areas in Russia. Thus studies on these landforms are crucial in understanding the carbon – permafrost feedback, and the warming climate scenario.
1.2 Previous research on palsas in Fennoscandia

The understanding of palsa distribution and formation in Fennoscandia has improved during the last decades, however palsas and peat plateaus are complex features and their formation and degradation depend on environmental factors which are still not fully comprehended (Borge and Westermann et al., 2016). But with an accelerated distribution of in situ measurements during the IPY (Christiansen and Etzelmuller et al., 2010, Juliussen and Christiansen et al., 2010, Romanovsky and Smith et al., 2010, Hipp and Etzelmüller et al., 2012), modeling of permafrost have become a powerful tool in investigating ground temperatures. However, as permafrost and frost related geomorphological processes are sensitive to climate change, and are controlled on topography and ground properties, a demand for topographic parameterization in permafrost modeling have been expressed (Isaksen and Hauck et al., 2002, Etzelmüller, 2013).

Studies on permafrost and the ground thermal regime of permafrost in Norway have been performed by numerous of authors through the last decades (e.g. Åhman, 1977, Juliussen and Humlum, 2007, Farbrot and Isaksen et al., 2013, Gisnås and Etzelmüller et al., 2016), which have all concluded that ground temperatures are increasing. Studies on permafrost in mires have proven that palsas and peat plateaus in Norway are degrading (Sollid Sørbel, 1998, Etzelmüller and Hoelzle et al., 2001, Sollid and Isaksen et al., 2003, Borge and Westermann et al., 2016), and the situation in the neighboring countries, Sweden and Finland, is no different (Fries Bergstrom, 1910, Salmi, 1970, Luoto and Fronzek et al., 2004c, Zuidhoff Kolstrup, 2005, Seppälä, 2011). This is indicated by increasing of ground temperatures (Gisnås and Etzelmüller et al., 2013, Farbrot and Isaksen et al., 2013, Westermann and Schuler et al., 2013), and thus a thickening of the active layer depth (Zuidhoff, 2002, Åkerman and Johansson, 2008), following an extensive degradation in lateral extent (Borge and Westermann et al., 2016) which may be leading to the increased appearance of thermokarst lakes (Matthews and Dahl et al., 1997, Luoto Seppälä, 2003) in continuos permafrost areas, and decreased areal extent of thermokarst lakes in discontinuous permafrost areas(Luoto Seppälä, 2003, Kirpotin and Polishchuk et al., 2009).
The regional distribution of permafrost in Norway is relatively well known, from field investigations (Etzelmüller and Hoelzle et al., 2001, Sollid and Isaksen et al., 2003, Farbrot and Isaksen et al., 2008, Isaksen and Farbrot et al., 2008, Isaksen and Ødegård et al., 2011, Hipp and Etzelmüller et al., 2012, Farbrot and Isaksen et al., 2013), from bottom temperature of winter snow cover surveys (Isaksen and Hauck et al., 2002, Heggem and Juliussen et al., 2005, Gisnås and Westermann et al., 2015, Gisnås and Schuler et al., 2016), landform inventories (Soillid and Sørbel, 1998, Etzelmüller and Hoelzle et al., 2001, Lilleøren and Etzemüller, 2011, Borge and Westermann et al., 2016), generalized linear modeling (Borge, 2015), equilibrium permafrost modeling (Juliussen Humlum, 2007, Gisnås and Etzelmüller et al., 2013) and by transient thermal modeling (Westermann and Schuler et al., 2013).

Generally in Southern Norway the lower limit of discontinuous mountain permafrost decreases in elevation from West (approximately 1600 m.a.s.l. to East (approximately 1200 m.a.s.l.), here permafrost is mainly restricted to areas above the tree line and sporadic occurrences in palsa mires which exists below 1000 m.a.s.l. in Southern Norway (Sollid and Sørbel, 1998). In Northern Norway the lower altitudinal limit of permafrost is decreasing with the increasing continental climate away from the coast (Gisnås and Schuler et al., 2016) however it is also common near sea level in large mires with thick peat deposits (Farbot and Isaksen et al., 2013, Borge and Westermann et al., 2016). According to Farbrot and Isaksen et al., (2013) permafrost in Finnmark exists above c. 1250 m a.s.l; in contrast to the continental areas of Fennoscandia where permafrost is abundant down to 800-850 m a.s.l. (e.g. in Finland ((Ridefelt and Etzelmüller et al., 2008)) and to southeastern parts of Finnmark (Figure 3) where permafrost is widespread above the treeline (> 400 m a.s.l.).
1.3 Objectives

This thesis is based on a strong motivation for improving the soil and snow forcing in permafrost modeling, which are significant parameters for permafrost distribution and crucial factors for estimating ground temperatures in transient thermal modeling. As mentioned, climate is warming and leading to permafrost is degradation. Palsas demarcate the southernmost edges of discontinuous permafrost and are therefore good indicators on permafrost degradation in other colder regions in the future. Permafrost in mires contains immense amounts of organic material and consequently stores excessive amounts of greenhouse gasses. Henceforth, with the reported palsa and peat plateau degradation in the northern hemisphere, knowledge on climate – carbon feedback is required. This may be obtained by modeling the future ground temperatures for permafrost areas. However, permafrost models are often of coarse resolution, and forced with vegetation, soil and surface parameters which are obtained by datasets of a coarse resolution. From this, permafrost models may yield inaccurate ground temperatures, and skewness in the predicted permafrost scenarios.

However, the environmental factors governing palsas formation and degradation are not fully understood yet, and in order to enhance the knowledge on the subject, further investigation on mire stratigraphy and snow cover have been advised during the last decades. Therefore, this thesis has the main objective to enhance the model performance for permafrost in organic rich soil, by estimating some of the local factors controlling palsa formation. This would further enlighten the permafrost models and in a best case scenario yield a dataset which may be utilized in predicting the outcome of climate warming. From this, the objectives for this thesis are listed as follows:

- Infer the appropriate soil stratigraphy and winter snow cover governing in palsa mires in Finnmark, northern Norway.
- Determine the importance of local environmental and physical characteristics controlling palsa formation and active layer thickness Finnmark.
- Infer the temperature regime of the ground in permafrost areas in northern Norway, by utilizing a transient thermal model.
• Compare the modeled ground temperatures and active layer depths with a recent map of palsa distribution in the region.

• Relate the distribution of the current extent of permafrost in northern Norway to a future climatic scenario together with topography and land cover.
1.4 Thesis structure

The first part of Chapter 2 gives broader overview of permafrost in relation to climate and a general theory of palsas and their formation and response to climate change. The theory of transient thermal modeling of permafrost is presented in section 2.3, giving some of the physics behind the numerical model CryoGrid 2, which is the model used to infer the temperature regime of the ground in Finnmark.

The Third Chapter presents the area of study and gives the reader an overview of the climatic scenario in Finnmark as well as an overview of the geology and geomorphology of the county and the five study sites capsuled in the model.

All of the methodologies utilized in this thesis are presented in Chapter 4. Firstly the methods for the fieldwork are presented (4.1), following the methods for the model operation (4.2) and the forcing data used in the initialization (4.3).

All of the results for this thesis are described in Chapter 5, and divided into field work observations (4.1) and model results on a local scale for the five study sites (5.2) and a regional scale for Northern Norway (Finnmark) (5.3).

The results are discussed in Chapter 6 in a similar structure as in Chapter 5. A conceptual model for palsa degradation based on fieldwork and Fourier’s Law of Heat Flow is also presented (6.2) plus a comparative analysis of the palsa distribution of Finnmark in relation to the regional temperature regime produced by CryoGrid 2 is also given (6.3). This Chapter is terminated by suggestions for further research in relation to the outcomes from this thesis.

Conclusions and highlights from this thesis are introduced in a minor paragraph, and summed up as bullet points in Chapter 7. The references used for this thesis is listed in the Literature section.

Results and figures that have importance for this thesis but which was not included in chapter 5 is presented in the Appendix, together with the Matlab ® script of CryoGrid 2, as was provided by Sebastian Westermann on the startup of this project in August 2015.
2 Theoretical Background

2.1 Permafrost and climate

As a consequence of global warming, permafrost is estimated to undergo a significant degradation towards 2100 (Delisle, 2007, Lawrence and Slater et al, 2008, Hipp and Etzelmüller et al., 2012, Gisnås and Etzelmüller et al., 2013, IPCC, 2014). Degrading is not only projected for areal extent, but also in deepening of the active layer (Anisimov and Shiklomanov et al., 1997, Christiansen and Etzelmüller et al., 2010) Recent research reveals that ground temperature measurements in boreholes in permafrost areas correspond to the projected situation, (Lemke and Ren et al., 2007, Christiansen and Etzelmüller et al., 2010, Romanovsky and Drozdov et al., 2010, Romanovsky and Smith et al., 2010)]. Borehole distributions have expanded during the International Polar Year (Christiansen and Etzelmüller et al., 2010). However, in situ ground temperature measurements still remain both sparse and costly, not to mention time consuming in maintenance. Thus, such in situ measurements remain somewhat limited. From this, extrapolating borehole temperature over large unmanaged areas may lead to uncertainties, especially in areas where there is a demand for a large number of boreholes due to strong climate –and height gradients and or strong heterogeneity of surface- and vegetation (Etzelmüller and Ødegård et al., 2001, Etzelmüller Frauenfelder, 2009, Westermann and Schuler et al., 2013]). However, another method to describe the vertical distribution of ground temperatures is by usage of thermal permafrost models (Goodrich, 1978; 1982, Zhang and Chen et al., 2003, Jafarov and Marchenko et al., 2012). Such powerful models are “forced” with time series of meteorological factors that affect the ground, such as snow depth and air temperature. These climate variables are obtained in different ways, e.g. through meteorological stations or atmospheric circulation models. Implementing dataset from these variables will enable modeling of permafrost temperature, and correlate this to point measurements in boreholes. Permafrost modeling yields insight in the future climate scenario and enlightens the possible consequences on the measured permafrost degradation. Thus research in estimation on the parameterization and classification of the terrain in permafrost region is crucial in understanding the permafrost – climate relation. When establishing an
empirical physical relationship between a topographic parameter (e.g. snow depth) and landforms one may assume that the observed relationship is scale independent (Etzelmüller and Ødegård et al., 2001), thus one may further use the topography–landform relationship in both micro scale permafrost modeling on e.g. one specific palsa mire, and further infer the observed connections and parameters to macro scale regional modeling. This thesis forms its basis on this theory, as in situ soil stratigraphy and snow measurements are presented and related to one another, as clear trends in soil stratigraphy and snow depth have been found on palsa mires in Finnmark (e.g. Figure 14Figure 19Figure 21Figure 26Figure 28

2.2 Climate-Permafrost System
According to Lachenbruch and Cladouhos et al. (1988) (reference therein Henry and Smith (2001)) the climate-permafrost relation can be represented schematically by the mean annual temperature regimes at three levels in the surface boundary layer. This is illustrated in (Figure 1). This Figure illustrates the surface offset, thermal offset and the geothermal gradient which are of importance in understanding the permafrost-climate relations. The surface offset defines the difference between the mean annual air temperature (MAAT) and the temperature at the ground surface (MAGST). Surface offset is a result of the vegetation on the summer and the winter snow cover (Smith Riseborough, 2002). A high snow cover will yield a higher surface offset temperature. The vegetation cover plays an important role in the surface temperature depending on vegetation type – and moisture. Dense vegetation during the summer will lead to less radiation reaching the ground surface leading to a depressing of the temperature. This is called the vegetation offset (Smith Riseborough, 2002).

During winter snow cover, the vegetation control both the persistence and depth of the snow (Smith Riseborough, 2002, Jean Payette, 2014). The MAGST is strongly controlled by the winter snow cover due to the low thermal conductivity of the snow which is resulting in the snow having an insulating effect on the ground as neither the cold air temperatures are reaching the ground and the loss of heat from the ground is also restricted during the coldest parts of the year. As a result, the ground surface temperatures (GST) are higher than the air temperatures during winter snow cover (Gold Lachenbruch, 1973). This is defined as the nival offset (Smith Riseborough, 2002). The annual effect of the nival – and vegetation offset yields the surface offset, and it is
often positive due to the stated warming effect of the snow on the ground surface (Smith Riseborough, 2002).

Generally the temperature in each level in Figure 1 will differ on an annual mean basis (Smith and Riseborough, 2002). And the temperature may be decreasing with depth. This temperature decrease in the soil defines the thermal offset, which is the difference between thee MAGST and the temperature at the top of permafrost (TTOP). The thermal offset is also a result of a difference in conductivity of summer and winter; however it is regarding the conductivity in water-rich soil. Liquid water has a high thermal conductivity, but the thermal conductivity of water in solid phase is four times as high (Goodrich, 1978, Smith Riseborough, 2002). This leads to the possibility of permafrost conditions for water-rich sediments, peat or soils that have a MAGST > 0 °C. The difference between summer conductivity and winter conductivity is defined as the conductivity ratio. Conductivity ratio may vary depending on soil type, as e.g. bedrock usually have a conductivity ratio of 1 while organic soils have a great range in conductivity ratio (0.3-1) from saturated to dry conditions (Smith and Risborough, 2002). Thus bedrock usually does not have a thermal offset, while organic rich sols have a large thermal offset.

According to Van Everdingen (1998) the depth of zero annual amplitude (DZZA), the depth at which the temperature does not have an annual fluctuations often define the mean annual ground temperature (MAGT). However, below this depth the temperature gradient is defined as the geothermal gradient (Figure 1). One commonly used permafrost temperature definition is the MAGT at the DZAA (10-20m depth), according to Henry and Smith (2001) this will normally be 0.2-0.4 °C higher than TTOP, because of the geothermal gradient. For northern Fennoscandia the DZAA in organic rich peat plateau is measured to be close to 0 °C, by monitoring borehole temperatures (Christiansen and Etzelmüller et al., 2010).
Figure 1: The relation between air temperature and the temperature of permafrost, as illustrated by (Smith Riseborough, 2002) presenting a schematic mean annual temperature profile through the surface boundary layer.

2.3 Palsa and Peat plateaus

2.3.1 Definition

"What is a palsa?" Link Washburn asked this question in 1983, expressing a confusing variety of the term. Seppälä (1972) later explained the term in a morphological sense: “The term palsa was originally used by the Lapps and Northern Finns, and in their languages (Lappish and Finnish) means a hummock rising out of a bog with a core of ice.” Lundqvist (1969 p.208) presented another etymology defining palsas in the late 60’s as “mounds of peat and ice occurring on bogs in the subarctic region”. And later Richard Åhman (1977, p 144) argued that a palsa is a hillock with a frozen core formed by a build-up of segregated ice in mineral soil, or in peat, with a 1 m or less to 7 m cover of peat. Washburn (1983) argued that palsa is commonly a substantial element forming either a core or as a cover over a mineral core. He further implied the definition “Palsas
are peaty permafrost mounds ranging from c. 0.5 to c. 10 m in height and exceeding c. 2 m in average diameter, comprising (1) aggradation forms due to permafrost aggradation at an active layer permafrost contact zone, and (2) similar-appearing degradation forms due to disintegration of an extensive peat deposit.” According to van Everdingen (1998) palsas and peat plateaus are subarctic permafrost landforms in mires defined as “a peaty permafrost mound possessing a core of alternating layers of segregated ice and peat or mineral soil material” and as “a generally flat-topped expanse of peat, elevated above the general surface of a peatland, and containing segregated ice that may or may not extend downward into the underlying mineral soil”, respectively. Hence they are climatic morphological features, distributed in the discontinuous permafrost zone, defined below.

2.3.4 Physical properties of peat and palsa formation

Even though palsas are found in mires, the permafrost is restricted to the palsas themselves, and the formation is based on the physical properties of peat (Seppälä, 1986, 1988). The formation of permafrost in palsta mires is related to the humidity of the peat, as dry peat insulates well, however wet peat conducts heat better (Seppälä, 1986, 1988) and if the peat is frozen the conduction is even better (Kujala and Seppälä et al., 2008). Kujala and Seppälä et al. (2008) found that the thermal conductivity of frozen peat samples at -5°C to -12°C would be almost four times higher than the same peat samples when thawed at 0.5 °C to 20 °C. The conductivity values for frozen peat \( K_{\text{frozen}} \) ranged from 0.41 – 1.48 Wm\(^{-1}\)K\(^{-1}\), while for the thawed peat \( K_{\text{thawed}} \) were in the ranges of 0.20 – 0.52 Wm\(^{-1}\)K\(^{-1}\), depending on the thaw temperature, soil samples and water content. During wintertime, the heat is extracted from deeper layers, and the winter cold will penetrate deep into the peat layers if the snow cover is thin (Goodrich, 1982, Seppälä, 1990, 1994, Zhang and Wang et al., 2008, Gisnås and Westermann et al., 2015). But during the summer, when the peat is dry, it will insulate the frozen core in the palsta, preventing the warm summer temperatures to infiltrate into the peat and reaching the ground ice. This leads to preservation of permafrost, and further enhancing of the ground ice within the palsta. According to Seppälä and Kujala et al. (2009), if the mean annual air temperature is close to -2 °C and the soil does not have an insulating layer of peat, the permafrost would disappear in palsta regions.
2.3.2 Palsa environment and distribution


However, Fennoscandian palsas are indicating the westernmost extent of a much larger peat plateau extent in the Russian tundra (Kirpotin and Polishchuk et al., 2009, Romanovsky and Droz dov et al., 2010, Kirpotin and Polishchuk et al., 2011), and the thoroughly reported degradation of palsas in the northern hemisphere (Laberge Payette, 1995, Zuidhoff Kolstrup, 2000, Luoto and Heikkinen et al., 2004b, Tremblay and Bhiry et al., 2014, Borge and Westermann et al., 2016) must be a window towards the future for the vast permafrost in Siberia. Normally, palsas demarcate the outer limit for permafrost in a given area as they are located within the sporadic permafrost zone (Sollid and Sørbel, 1998) and are found in a narrow climatic envelope (Parviainen Luoto, 2007). The permafrost temperature in palsas is thus relatively warm, with a mean
annual ground temperature (MAGT) often close to 0 °C in Northern Fennoscandia (Westin and Zuidhoff, 2001, Etzelmüller and Frauenfelder, 2009, Christiansen and Etzelmüller et al., 2010, Johansson and Åkerman et al., 2011). and according to Åhman (1988, p. 143) the MAAT is no higher than 0°C and often lower than that. Seppälä (2011) claims palsas are characteristic on mires in the southernmost zone of discontinuous permafrost in the circumpolar area of the northern hemisphere. Therefore, the palsas are very reliable indicators of changing environmental conditions, especially since climate warming have been predicted to be most extensive in polar regions (Anisimov and Shiklomanov et al., 1997) This type of environment is not existent in the Southern hemisphere due to a lack of adequately cold and dry land. For instance mires in Tierra del Fuego on the southern tip of South America do not contain permafrost (Seppälä, 2011). Suggestions of further investigations on palsas have thus been proposed (Seppälä and Kujala, 2009, Sollid and Sørbel, 1998, 2003, Christiansen and Etzelmüller et al., 2010, Westermann and Schuler, 2011, Borge and Westermann et al., 2016).

What are the characteristic climatic parameters in palsa regions? A question difficult to answer due to limited meteorological series available for the regions and no long-term climatic observations of the palsa mires. Further, the local climate patterns are formed by a combination of several climatic and atmospheric parameters such as: temperature, precipitation, wind, humidity, cloud cover and in and out-going radiation (Seppälä, 2011). Additionally topography, aspect and soil properties influence local ground surface climatic conditions, which affect the palsa behavior. The conditions required for palsa formation may be interpreted and hypothesized if long-term and detailed observations are available and processed using adequately statistical approaches. This has been done in e.g. northern Norway, Finnmark (Borge, 2015) which resulted in a report on a massive degradation in palsa mire area (-33 to -78 % ) during the last 60 years. Luoto and Hjort (2004a) concluded the optimum areas for palsa to have a low annual precipitation (<450 mm) and a MAAT in the range of -3°C to 5 °C. A spatial analysis of the distribution of palsa mires in Finnmark (Borge, and Solheim et al., 2016) concluded that palsa mires in this area are concentrated at a temperature range between -2.5 °C to -4 °C.
2.3.5 Palsas cyclic development

According to Seppälä and Kujala et al. (2009) frozen peat does not form segregated ice lenses by frost heave processes. And almost no segregation ice is formed in most peat types during freezing (An and Allard, 1995 p. 236). This means that the ice lens, which is present in palsas, is not normal ice segregation as one per example may observe in freezing silts. The palsa ice lens is rather formed by a saturated layer which is formed at the thawing front as the active layer in a palsa thaws and some water migrates into the frozen layer and eventually into the permafrost underneath (Smith, 1985). Field observations of new palsa embryos by (Seppälä, 1982, 2006) reveal that during summer the palsa will rise above the water saturated peat when the surrounding seasonal frost layer has thawed. As the water-saturated peat under the ground ice is being filled in with water by the surrounding wet peat, a void space will form under the upheaved frozen core. During the following winter, the water under the ground ice will freeze as the warmth is extracted from the deeper frozen core, and thus a new layer of ice is formed. This process will be repeated in succeeding summer as the thicker ground ice is uplifted by buoyancy, and the palsa will upsurge further above the mire surface and a new layer of water will migrate from the thawing active layer towards the bottom of the frozen core (Figure 2). A simple calculation of the possible uplift force, based on Archimedean law revealed that with a 1 m frost depth and water content of saturated peat was 80 % and its density 1100kgm\(^{-3}\) the uplift pressure would lift up the active layer of 18 cm (Seppälä and Kujala, 2009).

In Figure 2, the formation of palsa is illustrated, as in Seppälä and Kujala (2009). Palsa formation begins when the winter frost may infiltrate deep enough to prevent the summer heat from thawing it completely, this may only occur when the snow layer is thin enough (Seppälä 1982, 1986, 1988). During the end of the first summer (Figure 2-b) the small embryo palsa rises above the surrounding thawed seasonal frost layer. According to Outcalt and Nelson (1984) this uplift is caused by buoyancy of the freezing core. This embryo palsa has pale sphagnum moss on its surface due to desiccation (Seppälä, 1986, 2003, 2006). Due to the change in wetness on the palsa surface, the vegetation changes accordingly (Seppälä, 1988, Van Vliet-Lanoe and Seppälä, 2002; Kuhry, 2008; Kokfelt and Reuss et al. 2010). The dry moss cover insulates the frozen
core from the warm summer temperatures, and a shallow active layer forms, ranging from 25-30 cm in Finnish Lappland (Seppälä and Kujala, 2009). The late summer upheaval leads to some void to develop under the frozen core, which is floating on water-saturated peat – or water (Seppälä and Kujala, 2009). During the autumn the peat becomes wet again due to the decreasing rate of evaporation.

Due to the mentioned properties of peat, a wetter peat layer will lead to further cooling of the peat and for frost to penetrate deeper in the ground. When the second winter arrives, the embryo palsa is lifted further above the ground and the mound is more exposed to wind. Following, the early snowfall will be transported away from the palsa summit to the lee side of the palsa, or further. Leaving the palsa with even less snow than the former winter, and even more exposed to the cold air temperatures. Frost penetrates deeper than the last winter, and during the next summer (Figure 2-d) the mount is even more upheaved, accelerating the drying of the surface peat by wind action and wind transportation away from the palsa summit during succeeding winter (Figure 2-e). When the active layer thaws during the summer, water will migrate downwards to the permafrost top and keep it saturated (Figure 2-f). As one may observe in Figure 2-d-f thin layers of ice is present within the frozen core of the palsa. Seppälä and Kujala (2009) suggest that they are formed by buoyancy of the frozen core and frost that sucks water from below the freezing front when the frozen core of palsa evolves downwards. The downward freezing of the palsa core will continue until it reaches the till or silt layers at the base of the mire. At this phase the palsa is at the mature stage of palsa formation (Figure 2-g). At this point, the palsa has already started fracturing on the surface, and consists of steep edges, which works as traps for snow during winter and leads to insulating of the peat from the winter air temperature. Collapsing of peat blocks into water ponds surrounding the mature palsa and abrasion can remove peat from the palsa surface (Seppälä, 2003). Once the degradation of the palsa has started, and further collapsing of the palsa edges occur the palsa is exposed to snow during winter, and summer thawing accordingly. The peat loses its insulating peat layer, and the frozen core will eventually thaw. Thermokarst will enhance the thawing of the old palsa, which in the end is leaving a water body behind in its collapsed stage (Figure 2-h). Water bodies, ponds or open pet surfaces without vegetation may be interpreted as old.
degraded palsas (Seppälä and Kujala, 2009). From such pools, a new palsa may evolve after a phase of peat formation. Hence palsas may have a cyclic formation.

In permafrost environments, the active layer is the only component of the soil where major changes in its physical properties may occur during each annual cycle. These variations may be changes in water- or ice content, thermal conductivity, density and mechanical properties. (Cosenza and Guerin et al., 2003, Lemke and Ren et al., 2007) These possible variations are of critical importance for many natural phenomena and processes, such as vegetation, hydrology and cryogenic processes. Consequences of alterations of such processes lead to restrictions on land use, agriculture and infrastructure (Anisimov and Shiklomanov et al., 1997, Christensen and Johansson et al., 2004, Bosiö and Johansson et al., 2012). Therefore, studies and knowledge of active layer development is important for, ecology and engineering in cold regions (Harris, 1986, Westermann and Schuler et al., 2013).
Figure 2: Principal phases of palsa formation, from Seppälä and Kujala (2009). "Formation of ice-rich layers (white) under the freezing core by buoyancy. Water movement is indicated with grey/blue arrows and upheaval by black arrows. Fracturing of active layer marked in (e) and (f) by black lines on the palsa surface, these fractures have increased in (g) which will increase the incoming heat flux from the air towards the palsa ground ice, as the insulating peat layer no longer covers the frozen core completely. The result from the degradation in (g) is presented in (h) as the collapsed stage. At this point the frozen core have thawed to a thermokarst lake."
2.4 Permafrost modeling

2.4.1 Transient thermal modeling

Thermal permafrost models based on conductive heat transfer in the soil and snow pack have proven to be powerful and efficient tools in investigating the vertical distribution in soil temperatures (e.g. Zhang and Chen et al., 2003, Jafarov and Marchenko et al., 2012, Westermann and Schuler et al., 2013). Such models are forced with temperature series of surface climatological variables, such as air temperature, snow height and snow density, and ground and soil properties. These variables can be obtained from numerous of sources, climate stations, in situ measurements, or atmospheric circulation models (Goodrich, 1978, Lawrence and Slater et al., 2008b, Westermann and Schuler et al., 2013). Gridded data sets of the mentioned variables enable the model to run, and the model results enables permafrost maps to be constructed and evaluated against point measurements in boreholes and field observations.

2.4.2 CryoGrid 2

CryoGrid2 is a transient thermal permafrost model, which is able to calculate the ground temperature consistent with heat transfer in the soil and in the snowpack. The model is based on the surface- or air temperature as well as snow depth and density for the potential permafrost area. This model is able to produce the transient response of ground temperatures to a changing climate (Westermann and Schuler et al., 2013). One-year data series with the mentioned climate variables gives enough input for the model to predict steady state ground temperature and active layer thickness for a study site with permafrost. Its model physics is comparable to other well-expended permafrost models, such as GIPL2 (Geophysical Institute Permafrost Lab 2) e.g. Jafarov and Marchenko et al. (2012).

The internal energy and temperature exchange in the ground is entirely based on the premises of Fourier Law of heat conduction, and the latent heat that is produced by soil freezing- or consumed by thawing. The model does not incorporate movement of water in liquid-or gaseous form in the ground, which sets the premises that soil water content may only be altered over time by phase change from melting or freezing. In spatially
distributed modeling the area of study will be decomposed in grid cells and the model variables and parameters are assumed constant. For grid cells significantly larger than the size of the vertical modeling domain, heat flow is only considered in the vertical direction for each grid cell. Hence, lateral heat flow between neighboring cells will be neglected (Westermann and Schuler et al., 2013). In the following sections (2.3.3 and 2.3.4) the defining equations for a single grid cell are presented, as first imparted by Westermann and Schuler et al. (2013). Soil below the ground surface is regarded as positive z-coordinates, while the snow domain located above is regarded as negative z-coordinates.

### 2.4.3 Soil and Snow Thermal model

In the soil domain, temperature T changes over time t and depth z through heat conduction as only a physical process, as described by Jury and Horton (2004).

\[
C_{\text{eff}}(z, T) \frac{\partial T}{\partial t} - \frac{\partial}{\partial z} \left( k(z, T) \frac{\partial T}{\partial z} \right) = 0, \quad (1)
\]

where \( k(z, T) \) (Wm\(^{-1}\)K\(^{-1}\)) denotes the thermal conductivity and \( C_{\text{eff}} \) (Jm\(^{-3}\)K\(^{-1}\)) the effective volumetric heat capacity, accounting for the latent heat of freezing and melting of water and ice as following:

\[
c_{\text{eff}} = \left( c(t, z) + L \frac{\partial \theta_w}{\partial T} \right), \quad (2)
\]

where \( \theta_w \) denotes the volumetric water content, and \( L = 334\text{MJ m}^{-3} \) the specific volumetric latent heat of fusion of water. The first term is calculated from the volumetric fractions of the constituents as follows:

\[
c(T, z) = \sum_\alpha \theta_\alpha(T, z)c_\alpha, \quad (3)
\]

where \( \theta_\alpha \) and \( c_\alpha \) (Jm\(^{-3}\)K\(^{-1}\)) represent the volumetric contents and the specific volumetric heat capacities (following Hillel (1982) of the constituents water, ice, air, mineral and organic, \( \alpha = w, i, a, m, o \), as from (Cosenza and Guerin et al., 2003):

\[
k = \left( \sum_\alpha \theta_\alpha \sqrt{k_\alpha} \right)^2. \quad (4)
\]

The temperature-dependence of the thermal conductivity, which gives rise to the thermal offset between ground surface temperature and ground temperatures
(Osterkamp Romanovsky, 1999), is hereby contained in the temperature-dependent water and ice contents as detailed in section 2.3.4. The parameterization according to Equation (4) was chosen by Westermann and Schuler et al. (2013) for simplicity. This is due to insufficient reliable validation studies and thus a lack of standardization for a particular conductivity model so far.

Heat conduction is the only process of energy transfer in the snow domain in CryoGrid 2, similar to Equation (2). The snowpack is considered “dry”, i.e., production, infiltration and refreezing of melt (and rain) water are not considered. Thus, the density of the snowpack is constant. The thermal conductivity and volumetric heat capacity of the snowpack, $k_{\text{snow}}$ and $c_{\text{snow}}$, are assumed to be constant in time and space. The volumetric heat capacity is calculated from the densities $\rho$ of snow and ice as follows:

$$c_{\text{snow}} = \frac{\rho_{\text{snow}}}{\rho_{\text{ice}}} c_{i}.$$  

(5)

### 2.4.4 Soil freezing characteristics

According to Westermann and Schuler et al., (2013) Dall’Amico and Endrizzi et al. (2011) sets the basis for the soil freezing characteristics $\theta_w(T)$ in CryoGrid2, relating soil freezing to its hydraulic properties assuming that freezing has the same effect on the soil water potential as drying (see also Daanen et al., 2008 and Daanen and Nieber, 2009). The soil water retention curve is parameterized as following according to Mualem (1976) and Van Genuchten (1980) (reference therein, Westermann and Schuler et al., 2013): the soil water potential in unfrozen state, $\psi_{\text{w0}}$ (m), is expressed as a function of the volumetric soil water content, $\theta_{\text{w,m}}$, the saturated volumetric soil water content, $\theta_{\text{ws}}$, (corresponding to one minus soil porosity) and the residual volumetric soil water content in frozen state, $\theta_{\text{wr}} = \theta_w(T \ll 0 \, ^{\circ}\text{C})$ as:

$$\psi_{\text{w0}} = -\frac{1}{\alpha} \left[ \left( \frac{\theta_{\text{w,m}} - \theta_{\text{wr}}}{\theta_{\text{ws}} - \theta_{\text{wr}}} \right)^{\frac{n}{1-n}} - 1 \right]^{\frac{1}{n}},$$  

(6)

where $\alpha (\text{m}^{-1})$ and $n (-)$ are parameters related to the soil type. The residual volumetric soil water content in frozen state becomes zero for saturated conditions. For unsaturated conditions, the temperature $T^*$ below which soil water starts to freeze, is depressed below the freezing point of free water $T_f = 0 \, ^{\circ}\text{C}$, as:
\[ T^* = T_f + \frac{g T_f}{L} \psi_{w0}, \]

where \( g = 9.81 \text{ms}^{-2} \) denotes the standard acceleration of gravity. Accordingly, the soil water potential, \( \psi(T) \), decreases linearly with temperature for temperatures below \( T^* \):

\[
\psi(T) = \begin{cases} 
\psi_{w0} + \frac{L}{g T_f} (T - T^*) & \text{for } T \leq T^* \\
\psi_{w0} & \text{for } T > T^*.
\end{cases}
\]

A detailed derivation of Equations (7) and (8) is provided by Dall’Amico and Endrizzi et al. (2011). Employing the temperature dependent soil water potential \( \psi(T) \), see Equation (8), the soil freezing characteristics \( \theta_w(T) \) is again obtained by the Mualem-van Genucgthen model as following (Westermann and Schuler et al., 2013):

\[
\theta_w(T) = \theta_{wr} + (\theta_{ws} - \theta_{wr})(1 + [-\alpha \psi(T)]^n)^{1 - \frac{1}{n}}.
\]

Note that \( \theta_w(T) \) is fully defined for given values of \( \theta_{ws}, \theta_{wr} \) and \( \theta_w(T > 0 \text{°C}) \) through Equations (6) to (9). The volumetric ice content is hence given by:

\[
\theta_i(T) = \theta_{ws} - \theta_w(T).
\]

In CryoGrid2, two soil types have been distinguished sand and silt, for which the values of \( \alpha, n \) and \( \theta_{wr} \) have been chosen according to Dall’Amico and Endrizzi et al. (2011).


3 Area of Study

3.1 Geographical Setting

3.1.1 Study Area

This thesis concentrates on North-South and East-West transect of areas of possible palsa distribution in Finnmark County, Northern Norway (69-71° N and 21-32° E), which follows the recent palsa mire mapping by Borge (2015) and Borge and Westermann et al. (2016). However, this thesis also includes the southernmost tip of Finnmark, Aidejavri and one of the most extensive palsa mires in Finnmark, located near Karlebotn have also been investigated. Hence, the west–east transect contain Lakselv in the innermost part of Porsangerfjorden and Karlebotn in the innermost part of Varangerfjorden, and north–south transect comprise Lakselv in the northwest to Karasjok and Kautokeino in Southernmost part of Finnmarksvidda.

Finnmark is the largest county of Norway, covering an area of 48,618 km² (SNL, 2015) with an increasing in both elevation from East to West, and in temperature gradient from south to North, and according to senorge.no (Engeset and Tveito et al., 2004a, 2004b, Mohr, 2008, Mohr Tveito, 2008) a decreasing in snow depth from the continental areas to the coast (see Figure 50). Furthermore, Finnmark County is located within the Arctic Circle and borders to Troms County in the west, Finland in Southwest and Russia in the North-Easternmost part. Finnmarksvidda, located in the southeastern part of Finnmark County, is an undulating peneplain, mostly constricted to 300-500 m a.s.l. In fact (95%) of the county Finnmark is situated below 600 m a.s.l (Farbrot and Isaksen et al., 2013). The highest elevations in the County (up to 1200 m a.s.l.) occur in the Southern most parts next to the border of Troms County. However, the main mountain range is the Gaissane Mountains, which has peaks up to 1100 m a.s.l. In the extreme northeast of Norway, Varangerhalvøya is situated, with mountain plateaus dominating with elevations up to 600 m a.s.l.
3.1.2 Study Sites

Figure 3 illustrates the location of Finnmark in Northern Fennoscandia in the inlet map, and the four study sites implemented in this thesis. The meteorological stations for the study sites are also implemented.

Figure 3: Presenting the study area, Finnmark, in Northern Fennoscandia (inlet map to the bottom right) and the two coastal study sites (Lakselv and Karlebotn) and the three continental sites (Iskorás, Soussjavri and Aidejavri) and the adjacent meteorological stations. Inlet map has is projected in ETRS 89 Geographic Degrees. See legend.

The palsa mire in Lakselv, Oksebergmyra, is located right on the coastal line and located at (EUREF89 UTM 35) 7110625.19N, 877841.71E, almost at sea level, 25 m a.s.l.
Oksebergmyra covers an area of 456133 m$^2$ and 250 meters from the mire in eastern and western direction; mounts up to 60 meters enclose the palsa mire. Across the fjord there are summits up to 550 m a.s.l. The mire is situated on a belt of sandstone (quartzite) and is under the marine limit yielding great likeliness of a marine clay cover being present in the mire (Engevik, 2010, NGU, 2012). Thick marine deposits range 200 meters south of the mire.

Karlebotn is also a coastal study sites with palsa mires located almost at sea level near the innermost part of Varangerfjorden in Loakhejeaggi (EUREF 89 UTM 35) 10006825.81E 7835430.17N. This palsa mire covers more than 5 km$^2$ and several palsa mires are located in this area. The mires itself situated on conglomerates and sedimentary breccia. This area was also under the marine limit after the last glaciation, and hence NGU (2012) and Høgaas and Hansen (2010) classify this area to have marine clay deposits. The mire is surrounded by oceanic / fjord deposits, river deposits and bare mounts of gneiss and granite are surrounding this mire in the Southern end (Engvik, 2010, NGU, 2015), ranging up to 210 m a.s.l.

Two separate adjacent palsa mires north of Iskorás Mountain have been studied for this thesis. Lindejaeggi is situated 3 km north of the mountain ridge Iskorás, while the western palsa mire is 2 km northwest from the mountain. The study sites are about 10 km north of Karasjok, in the easternmost part of in Finnmarksvidda (Figure 3). A thick cover of moraine surrounds the mires, and the bedrock under the moraine consists of gneisses and mica. The Iskorás Mountain is mapped to be of quartzite (Engvik, 2010, NGU, 2015). The study sites are located 360 m a.s.l, and thus above the marine limit after the last glaciation.

Soussjavri palsa mire is located 50 km southwest from Karasjok at 310 m a.s.l and covers an area of 0.9 km$^2$. A thick cover of moraine surrounds this mire as well, with elements of other glacial deposits, e.g. glacial fluvial deposits from the minor valley southwest of the study site. These sediments terminate in the palsa mire. Lake deposits from the Soussjavri lake is deposited in the Northern perimeter of the mire. The bedrock of Soussjavri consists of granitic gneisses.

The Southernmost study area is Aidejavri palsa mire, located 30 km southeast from Kautokeino, near the border of Finland. A thick cover of moraine and glacial fluvial
deposits also surrounds Aidejavri, while the bedrock is made up of mafic greenstone and amphibolite. The elevation of the palsa mire is 370 m a.s.l.
3.2 Geomorphology and Geology

The area of study, Finnmark, is of somewhat varying landscape, as mountains and fjords dominate the western coast, and further east towards Finnmarksvidda the scenery is rather different with a smoother landscape of round mountains and the Finnmarksvidda. Thus, the whole county is, just as the rest of Norway quite influenced by the early glaciations in Pleistocene Epoch (2.6 Ma to 11.7 thousand years ago) (Nordgulen and Andersen et al., 2008) One may observe the glaciers influence as macro scale features e.g. in the North and West where long and wide fjords dominate, meso scale forms such as moraine types of e.g. long eskers, flutes and drumlins and micro scaled forms such as bedrock scouring may also be observed all over Finnmark. As mentioned in section 3.1 Finnmark have large glacial deposits consisting of moraines and glacial fluvial sediments. In the lower lying coastal areas marine deposits yields opportunities for agricultural land.

The geology of Finnmark is somewhat from different time scales and bedrocks from southeast to northwest. The continental areas make up the oldest bedrocks in the area from Archea and Proterozoicum (3500 – 2500 Ma), while the coastal belt consists of nappes from Caledonian Orogenese (750 – 400 Ma) (Nordgulen and Andresen et al., 2008) Hence, the coastal areas are of metamorphic sedimentary rocks while the continental bedrock consists of tonalitic gneisses in the center of Finnmarksvidda which is surrounded by a belt of mineral rich greenstones (Nordgulen and Andersen et al., 2008).
3.3 Climate

Northern Norway is influenced by the warm North Atlantic Current (NAC), which leads to a strong temperature gradient from coastal areas from northwest to east in the continental parts of Finnmark (Johannessen, 1970). The NAC causes an ice-free coast at during the winter (Danevig, 2009). The coastal part of Finnmark is therefore relatively warm with a normal annual air temperature for the period 1971-1990 of 4 °C to 0 °C, see Figure 50. The summer temperatures are in the ranges of 10 °C to 12 °C on the coast, while the continental areas are warmer, up to 14 °C (Danevig, 2009). February is the coldest month on the coast (-2 – -7 °C), but the winters in the continental areas are much colder and even warmer than the coast during summer (Mohr, 2008, Mohr Tveito, 2008). Hence, the temperature range between summer and winter is more yields a continental climate in the innermost parts of Finnmark, and a coastal climate in in the northwest.

The normal annual air temperature - and precipitation on the continental parts of the county during the period 1971-1990 was -1 to -6 °C. However, most of Finnmarksvidda have a normal annual temperature of -2 to -4 °C, and a mean summer- and winter temperature of 8 to 10°C and -15 to -20 °C, respectively. It is the mountainous area of Gaissane that make up the coldest normal annual air temperatures, dropping below -5 to -6 °C. Nonetheless, Finnmarksvidda have been reported to have the coldest winter temperatures in all of Europe. And it is not un-normal for the temperature to drop below -40 °C during the winter months. Also, large inversions are measured in the same areas, for example during January 2016 , near Karasjok, Iskorás (591 m a.s.l) had a temperature of -17.4 °C while lower lying areas (131 m a.s.l), Márkannjárga, 19 km of Iskorás met station had -34.1 °C, and the minimum temperatures also differed with nearly 20 °C (met.no, 2016), reflecting the importance of temperature gradients with elevation.

Regarding precipitation, most coastal areas had a normal annual precipitation amount of 750 – 2000 mm during 1971-2000. In contrast, the continental parts are as much as four times drier and ranges from under 500 to 750 mm, respectively in the same period. The
normal maximum snow depth for 1971-2000 is decreasing with distance from the coast. Finnmarksvidda has an annual maximum snow depth within the ranges of 25-100 mm, while the coastal areas measured an annual maximum snow depth of 50 to > 400 mm, in the 30 year normal period.
4 Methodology

This chapter describes the different methodologies utilized in this thesis. The chapter is divided in two main sections: methods regarding field observations and (4.1) and methods regarding model operation (4.2). The structure of the model operation is encapsulated in a flowchart below, in Error! Reference source not found..

As Figure 4 illustrates, in order to obtain the ground temperatures in permafrost modeling, sufficient inputs and forcing data is required in order to reproduce the state of art, and future scenario of permafrost temperatures for the study sites. Climate forcing data as air temperature and precipitation is essential, and obtained from in situ meteorological stations. However for CryoGrid 2, one defines the snow depth by simply defining the snow depth build up and duration of snow cover, and the snow density. The air temperature and snow depth data sets are obtained from the database of senorge.no (Mohr, 2008, Mohr Tveito, 2008, Saloranta, 2012), which retain daily measurements on e.g. air temperature, wind, snow depth and precipitation interpolated between meteorological stations, and are available for download on senorge.no and eklima.no.

The climate forcing is described in section 4.2.1 and presented in section 5.1.1. Moreover, in situ snow depths and densities were also obtained during one week of field work in March, 2016, these measurements are presented in section 5.1.3, and have been used in the initialization of the transient thermal modeling of the selected study sites. Thus, sufficient datasets on snow forcing was available, and several snow scenarios have been used in the modeling of the palsa mires.

The soil stratigraphy data is also essential when modeling permafrost; however with a vast spatial variability of the topography, vegetation and soil stratigraphy, uncertainties are yielded in modeling ground temperatures over large areas. And with a spatial resolution of 1km², difficulties in accurate soil forcing apply. However, this thesis is based on several weeks of field work, whereof an interpretation on the soil stratigraphy in palsa mires in Finnmark have been conducted. The soil stratigraphy forcing is presented in section 5.1. CryoGrid 2 is based on physical laws of soil, surface and snow
interactions, which is presented in section 2.4 and thus will not be further debated in this chapter.

**Figure 4:** Illustrating the forcing data needed to CryoGrid 2 to run and produce ground temperatures. Stratigraphy and soil properties, in situ measurements of climate data and snow and temperature data from met stations.
4.1 Field work

On a general basis in several science disciplines the objective is often to prove field observations – and measurements to be coinciding with a theoretical background, and physical geography discipline is no exception. A rule of thumb for a geographer is to be able to explain physical phenomena in the field with acknowledged theory, or propose theories for observations. This is the key for science to evolve. In order to produce realistic model results, in this case from CryoGrid 2, one is in need of empirical data both regarding soil parameters and climate forcing.

During the last week of August until mid-September 2015, and one week during mid of March 2016 fieldwork in Finnmark and Troms was conducted. The objective of these field trips was to obtain a representative soil stratigraphy, and to get an overview of the snow distribution during the winter, for both palsas and wet mire sites in Finnmark. Study sites were chosen on a basis of possible permafrost and palsa sites, on the recent mapped palsa distribution (Borge and Westermann et al., 2013), present GST-loggers from summer 2014 and of course the relatively important factor of accessibility in a County with limited infrastructure.

The general observations have been of great value for the model initiation concerning palsa mire stratigraphy, insight in small-scale differences governing around palsa mires during different seasons, as well as gaining an overview of the study area and Northern Norway in general.

From Westermann and Schuler et al., (2013) it was clear that a proper stratigraphy for palsa mires was needed in order for CryoGrid2 to produce representative ground temperature for such permafrost regions. Hence, Lakselv, Iskorás, Soussjavri and Karlebotn mire sites was investigated during late August – early September 2015. Unfortunately, Aidejavri site was not investigated during summer due to limited time and resources. Soil samples were taken on palsas in Lakselv and Iskorás and wrapped in plastic and returned to the lab. The samples were described in the field, and the wetness of the soil was estimated visually and classified as either dry or wet. Organic content was analyzed in the lab; see Figure 11, Figure 12, Figure 13 and Figure 14. Wet mire sites were also examined, however, no soil samples were taken here due to the very high
wetness and the following uncertainties as the samples was not analyzed before late September in the lab. Nevertheless, one is able to observe the general soil properties in the uppermost parts of the soil with unsophisticated tests in the field. By simply squeezing the water out of a soil sample, one is able to conclude representative volumetric water content.

Lab work was performed quite some time after the field work, however since the samples were vacuum packed and was going to be dried before burning, hence the results should not be significantly skewed. The methodology for the lab work was performed as follows: Peat samples were firstly dried in plastic containers for 5 hours on 105-110 °C, and afterwards homogenized and grounded in a mortar before resampled into smaller ceramic containers. Containers were weighed without samples and thereafter weighed with the samples. The containers with the samples were subsequently burned at ± 400 °C for 24 hours. Samples were then weighed again; yielding the total mass lost on ignition (LOI), see Figure 12. From these observations dry bulk density (DBD) (Figure 13) and total volumetric content (Figure 14) of organics and mineral was calculated according to (Craig, 2004) and Skaven-Haug (1972) (reference therein: Andriesse (1988) p.155:160). From these observations, interpretations for a representative mire stratigraphy could be made. Hence, stratigraphy for a representative palsa, embryo palsa and wet mire stratigraphy is presented in Table 3. These soil stratigraphies were incorporated in CryoGrid 2, yielding characteristic mire stratigraphy in consistence with the realistic situation for a palsa mire in Finnmark.

Active layer measurements were performed during late summer 2015 in Iskorás, Soussjavri, Karlebotn and Lakselv. The measurements was performed along transects with programmed random points in the mire where GST-loggers was going to be installed. Every point where a GST-logger was installed, the AL was measured and soil wetness was noted. Real time kinematic (RTK) was used in the exact positioning of each point, yielding up to centimeter -level accuracy positioning of each point. The ALD was measured with a soil drill, which gives an approximate indicator of the thaw depth at the drilled point. Every point was drilled, a few centimeters from the GST-point, until reaching rock hard soil, which was interpreted to be frozen soil. On some points one was able to reach the actual ice lens (Figure 10 (2)). The resulting active layer depths are presented in section 5.2.2. In Iskorás, Soussjavri and Aidejavri orthophoto was
conducted by drone photography by equipment from University of Oslo (UiO), yielding very good resolution (10 cm) on the images (Figure 26, Figure 29, Figure 30, Figure 31, Figure 32). Karlebotn measurements were however performed on selected palsas and wet mire sites and are only presented by the average ALD in Figure 25. Karlebotn and Lakselv sites were not photographed with drone. However, the in situ snow measurements from Lakselv is given in the Appendix as a histogram plot with ALD, soil condition and snow depth (Figure 46).

Snow depth measurements were performed during the period of March 11th – March 15th 2016 on the same transects as the summer measurements in Lakselv, Iskorás, Soussjavri, and Aidejavri. In order to investigate the reliability of the model initialization of $\rho = 300 \text{kgm}^{-3}$, snow density measurements was also conducted on selected palsas and wet mire points in the four mentioned sites. Due to limited available time, Karlebotn was not investigated during this winter field trip.

By measuring weight of snow at vertical intervals (top, middle and bottom) on palsas, wet mire sites and forested areas, one is able to conclude on a reasonable snow density value for CryoGrid2 initialization. From equation 11 one is able to calculate the snow density (DeWalle Rango, 2008).

$$\rho = \frac{m}{V} \quad \text{ (11)}$$

Where $\rho$ equals snow density in $\text{kgm}^{-3}$, $m$ equals snow weight in kg and $V$ equals volume of the snow sample in $\text{m}^3$.

Due to the snow density measurements being conducted late in the fourth semester of this master degree, assumptions were made that snow density was according to earlier observations.

### 4.2 Model operation

In this thesis, the heat transfer equation (Equation 1) is numerically solved for a soil domain set to 100 m depth, using the method of lines (Schiesser, 1991, references therein Westermann and Schuler et al., 2013). Central finite differences are employed to discretize the spatial derivatives $\frac{\delta}{\delta z}$ on a grid, while the time $t$ is left as a continuous
variable. From this, derivatives are replaced by difference quotients, with temperatures assigned to the midpoint of a grid cell and fluxes to the boundary between two cells. The temperature of each grid cell is calculated according to Westermann and Schuler et al. (2013), and assigned to the midpoint of each cell, whereas the time is a continuous variable. And in order to account for large temperature gradients close to the surface, an increase of grid space, $\Delta z$, with depth is set according to Westermann and Schuler et al. (2013). The temperature within the grid cells containing snow is set to NaN, hence the snow temperature is not presented in this thesis.

### 4.2.1 Climate forcing data

Datasets with snow depth and air temperature is operationally provided by The Norwegian Meteorological Institute (Tveito and Førland, 1999, Mohr 2008, Mohr and Tveito, 2008,) and the Norwegian Water and Energy Directorate (Engeset and Tveito et al., 2004a,b, Saloranta, 2012) at a spatial resolution of 1 km². The data is available for download on senorge.no and eklima.no. Datasets of snow depth and air temperature have been chosen from the nearest met station for each study site. The met stations are Banak Airport Meteorological Station (St. nr. 95350) for Lakselv, Tana Bru (St. nr. 96850) for Karlebotn, Iskorás 11 (St. nr. 97710) and Karasjok (St. nr. 97251) for Iskorás site, for Soussjavri site data from the met station Couvddatmohkki (St. nr. 97350) was used and for Aidejavri the Kautokeino met station was used to collect the meteorological data. All of the data was downloaded from Senorge.no. The meteorological data used in the model is hereby denoted as Senorge data.

The Senorge dataset is based on interpolations of in situ measurements at meteorological stations distributed across Norway and implements the topographic through monthly varying lapse rates (Mohr and Tveito, 2008). Some assumptions are made in the interpolation and measurements: for the air temperature, the lapse rate varies as a function of altitude, average altitude of the surrounding area of the met station and the geographical location. The precipitation increases with 10 % per 100 m elevation increase, from 0 to 100 m a.s.l., and by 5 % per 100 m above 1000 m a.s.l. Assumptions that precipitation is in the form of snow if the air temperature is < 5 °C.
The snow depth is calculated from snow water equivalent (SWE), which is calculated from the precipitation and air temperatures (Engeset and Tveito et al., 2004 a,b).

The snow forcing is set to have a steady state density according to field observations, and does not change during the snow cover period. These assumptions are needed to be made, as snow density measurements were only conducted during a few days in March 2016. Sensitivity runs for different snow depths have been made and are presented in Chapter 5. Some simplifications are made in the model, e.g. snow pack is also assumed homogenous throughout the year. This was also the case when CryoGrid 2 was implemented for permafrost modeling in Southern Norway, and in situ measurements in this part of Norway was concluded to be more or less equal from top to bottom of the snow pack. From this, one may regard the assumption of a constant snow density with depth and time during the time of snow cover. However, large differences in snow density have also been reported (Lundberg and Richardson-Näslund et al., 2006), and the snow density will increase theoretically increase with depth due to compaction of the overburden pressure (Kojima, 1967), not to mention with persistence due to the compaction by wind. Nonetheless, a stable state snow pack with regard to snow density have been implemented on the basis of the field work during March 2016, and initialization before this field work were based on literature for snow density in northern Fennoscandia (Lundberg and Richardson-Näslund et al., 2006, Gisnås and Westermann et al., 2014, Gisnås and Westermann et al., 2015) The snow forcing scenarios are presented in Chapter 5.

Other factors needed to be parameterized in order for CryoGrid 2 to be run is the nFactor, which ratio between freezing degree days (DD) at the ground surface (s) and air (a) (Gisnås and Etzelmüller et al., 2013). The FDDs are defined by a temperature below freezing, freezing degree days (FDD) and temperature above freezing which is further defined as thawing degree days (TDD) This ratio yields a calculation on the surface offset and MAGST. Surface offset and MAGST may be calculated on a seasonal basis using freezing (nF) and thawing (nT) indices. The nF is thus the ratio between FDDs and FDDa (Gisnås and Etzelmüller et al., 2013). However, as ground temperatures were only conducted for one single palsa, and one wet mire point in Lindejaeggi near Iskorás, estimations on this parameter was in need. The nFactor was decided according to the
MAGT, e.g. when the temperature curve in the ground at approximately two meters depth was not stable with the forcing period and not varying from year to year. Thus some experimenting was performed in the first stages of the modeling. The empirically hypothesized nF was further compared to other studies on snow distribution modeling in permafrost areas.

4.2.2 Ground Properties

The forcing data regarding the soil properties are a result of the in situ observations from summer 2015, and some assumptions regarding the depth of mineral layer and bedrock is made according to Westermann and Schuler et al., (2013). Research on vegetation and ground properties on permafrost in organic rich soil have been encouraged to be investigated and parameterized in future permafrost modeling (Etzelmüller and Hoelzle et al., 2001, Etzelmüller and Ødegård et al., 2001). Recent transient thermal modeling of Southern Norway (Westermann and Schuler et al., 2013) have proven the inaccuracies in estimations on the soil stratigraphy by utilizing meso-scale geology and geomorphological maps, instead of systematical in situ observations for each study site. Thus in order to produce as realistic as possible scenario for the ground temperatures in permafrost areas, CryoGrid is forced with stratigraphy parameters obtained from systematical observations in palsa mires on the study sites, listed in Table 1Figure 5Figure 8Figure 9.

Generally Northern Norway, as stated in Chapter 3, is significantly affected by the last ice age with the surface sediments of thick moraine and tills covering severe parts of Finnmark. There have not been distinguished between different bedrocks or surface sediments in this thesis, besides the organic rich surface layer covering the mire surface.

The depth and location in the soil of the stratigraphic layers, other than the organic rich layer, are not mapped in this thesis, and assumptions are made and presented in Table 3. Therefore, the ground properties used as forcing data for CryoGrid 2 in this thesis is according to Westermann and Schuler et al., (2013). Nonetheless, for palsa mires and peat plateau modeling, it is the mineral content and organic content as well as the properties of these two contents that set the premises for permafrost formations as long
as the climatic conditions are favorable for permafrost formation. Thus, the soil forcing is regarded sufficient enough for model validation.

The thermal conductivity of the mineral component of the soil was set to 3 Wm\(^{-1}\)K\(^{-1}\). And the heat flux at the lower boundary was set to Wm\(^{-2}\). The model was run with Matlab 2014b\(^\circledast\) and the model results was analyzed and presented with Esri\(^\circledast\) Software ArcMap 10.3.1 as well as Excel 2013\(^\circledast\).
5 Results

This chapter includes meteorological data used in the initialization of the model (5.1.1), the observations from fieldwork during late summer 2015 and winter observations from March 2016 (5.1.2 – 5.1.3 and 5.2), from the model run concerning the five study sites (5.3) and regional model run concerning Finnmark (5.4). Section 5.1 contains time series of climatic forcing data from 2014-2015 used in the model initialization. The climate data is obtained from meteorological stations close to the study sites and from GST-loggers in one of the study sites. Section 5.2 presents active layer measurements and results from soil samples from the field, as well as some general observations regarding palsa mire degradation. It also includes field observations regarding depth, distribution- and density of the snow on selected study sites. Section 5.3 presents model results for the three mire stratigraphy classes presented in Table 1, concerning permafrost temperature and active layer depth with different snow scenarios and mire stratigraphy. The last section, 5.4, concludes the model initialization in a regional scale for Finnmarksvidda, and presents possible palsa occurrence with the forcing data as given in Sønorge and stratigraphy as observed in the field. A comparative analysis of the modeled palsa distribution with observed palsa distribution by aerial images is also given here (5.4.3).

5.1 Field Work Observations

5.1.1 Temperature and Snow data

Air and Ground Temperature
Figure 5 presents the downloaded air temperatures from the five study sites. The locations of the met stations are presented in Figure 3. Notice the differences between the winter temperatures for the two coastal sites and the three continental sites. Soussjavri and Aidejavri were almost 10 °C colder than Lakselv and Karlebotn.
Figure 5: Air temperature from mid-August 2014- mid August 2015. The meteorological station’s id number is presented in each subtitle.

Figure 6: The mean annual air temperature obtained from the air temperature forcing in Figure 5. The study sites in the legend correspond to the overlying column. Notice the clear temperature difference between the coast and the inland.
Snow depth- and duration
The snow depth and duration was somewhat varying for the 5 study sites (Figure 8), e.g. Lakselv has the shortest duration of snow cover as the first snow fall was in October and the ground was snow free already in the beginning of March. In comparison, Soussjavri and Iskorás had the first snow fall in late August. The Southernmost study site, Aidejavri, had the thickest snow cover of over 0.6 m. Lakselv also had a deep snow, however this did not persist for long.
Figure 8: Snow depth obtained from climate stations that were presented in Figure 3, for the winter of 2014 – 2015. Notice the different snow depths and persistence of snow for e.g. Lakselv and Aidejavri. These snow forcing has been utilized in palsa scenario and wet mire scenario runs of CryoGrid 2. Iskorás Downscaled snow is utilized for modeling a palsa scenario in section 5.3.1.

The Snow Scenarios

According to the winter snow measurements (Figure 21, and Figure 22) and the senorge snow data (Figure 8), there was made interpretations that a general palsa- and wet mire location in Finnmark were overlain by very different snow depths during the winter season. Therefore, model runs with palsa snow scenarios and wet mire snow scenarios have been conducted, presented in Figure 34, Figure 35, Figure 36, Figure 37, Figure 38, Figure 39, Figure 40, Figure 41, Figure 42. Embryo palsa scenario was also initialized with palsa snow scenario, according to the observed snow depth on the possible embryo palsa in Lakselv (Figure 24). Notice that a wet mire snow scenario includes six times more snow than of a palsa snow scenario, and the duration of the two are somewhat different as well. The palsa snow scenario has steady state 10 cm snow from January to May, while the wet mire snow scenario has a steady state 60 cm snow during mid-Decemberto May.
Figure 9: Presenting the depth and duration of the palsa snow scenario and wet mire snow scenarios, which are determined according to in situ measurements during winter 2016 and Senorge data.
5.1.2 Palsa Stratigraphy Observations

In order to initialize a realistic palsa stratigraphy in the model, several soil samples from palsa mires in Finnmark were collected and analyzed. By digging vertical profiles as deep as possible on several palsas, it is possible to propose the volumetric values of the different components in the active layer of a palsa. Such samples were conducted in Karlebotn and Lakselv during summer 2015. The samples in both Lakselv and Karlebotn mires were taken on palsas, which were affected by block erosion and depression with large fractures. The fracturing made it easier to collect the samples as a proper spade was not available in the field, however the moisture content may have been different as the peat surface was exposed to air. The samples from Lakselv were taken from a 1 m high palsa with active layer measured to be 0.4 m. In Karlebotn the palsas were generally quite large both in lateral and vertical direction, the sample were taken from a 6 m high and 30 m long palsa/dome palsa with an active layer depth measured to be 0.65 m. In both sites the samples were taken a couple diameters in the lateral direction and 2 and 5 diameters in the vertical direction in the fracture. From it is possible to observe several layers of burned peat in the successive peat layers, and an ice lens from the lower most part of the active layer in the Karlebotn palsa. The black peat layers were warmer than the lighter layers of peat, indicating different albedo conditions.
Figure 10: Palsa with block erosion and successive layers of burned peat in Karlebotn (1) with black layers indicating possible forest fire or burning organic soil. To the left (2), soil sample taken from the lowermost part of the active layer of a palsa in Karlebotn with ice lens. Palsa with severe cracks where the soil samples “Lakselv Top” and “Lakselv Bottom” were taken is presented in lower left (3) and (4).
Figure 11: Field samples from top layers of palsa in Lakselv (1 and 2) and Karlebotn (3 and 4). Notice the fine grained and soft texture from all four the samples.
Sample 1 was taken on a palsa in Lakselv, 10-20 cm down in the soil; sample 2 was taken below sample 1, at 50-60 cm depth. Sample 3 and 4 were taken on a fracturing palsa in Karlebotn (see Figure 18), 10-20 cm and 30-40 cm below the ground respectively. In contrast to the four organic samples, a sample from Iskorás was also compiled, however this sample was taken 100 meters from the Northern side of the Iskorás Mountain, on the transgressing area from the mountain foot towards the words, on what was interpreted to be an old glacier river plain. This Iskorás sample, sample 5, was taken on 45-50 cm depth.

Sample 2 to 4 were observed to be consisting of a mixture of brown soil with small fossil birch and moss fractures, while Sample 1 was of very light weighted and beige. Sample 5, the Iskorás sample was visually interpreted to be consisting of mainly fine grained sediments in a mixture of gravel and large grained sand. From the TOC test the samples were interpreted to consist of silt and sand, and classified as very clayey silt as over 20 % of the sample consisted of clay (Craig, 2004). From Figure 11 it is possible to see the fine grain texture of beige and orange-brown sediments from Lakselv palsa (marked 1 and 2) and Karlebotn palsa (marked 3 and 4).

From the field observations in the coastal mires of Lakselv and Karlebotn and the continental mires in Iskorás, Soussjavri and Aidejavri, it was relatively clear that the palsas were more or less similar to one another when considering the stratigraphic properties. This was in consistence with the lab results presented in Figure 12, Figure 13, and Figure 14. A representative active layer of a palsa had a small fraction of mineral content (5 %), presumably clayey silt- or very fine sand, in a mixture with 10-20 % organic materials. For all four samples, the majority consisted of organic material (Figure 12). Lakselv Top and Bottom samples had 98.4 % and 97.2 % organic, respectively. More or less the same results was obtained from the Karlebotn samples, as these Top and Bottom samples had 91.6 and 97.0% organics, respectively. However, the Iskorás sample had only ≈ 40.0 % organics.

The DBDs were somewhat varying (Figure 13), and Karlebotn was over 0.5 gcm⁻³ and thus had the highest BDB, while Lakselv Top sample only measured one fifth of Karlebotn sample, about 0.1 gcm⁻³.
Figure 12: Total organic content lost on ignition (LOI). Notice the difference between the high value for the 4 peat samples compared to the inorganic sample from Iskorás.

Figure 13: The dry bulk density for each of the five samples, varying from about 0.1 – 0.55 g cm$^{-3}$.

Figure 14: Total volumetric content of organic and mineral. Averages of the samples have been included in Table 1, e.g. a total of 15% organic and 5% mineral content was chosen.
In Lakselv and Karlebotn other interesting observations were made. In Lakselv, a peat mound with low mire vegetation of a distinct reddish color predicted as new palsas were detected, see Figure 15 and Figure 16. The peat was of significantly lighter color than the larger and degraded palsas. The mounds were approximately 0.2 – 0.5 m above the general mire height, whereas in comparison with the general palsa heights, which for the Lakselv palsas which were in the mature- and degrading stage ranged about 1 m above the general mire height, and in Karlebotn the mature and degrading palsas/peat plateau are as high as 5-8 m. ALD was measured to 35 cm for the Lakselv palsa, while for Karlebotn the ALD was a bit higher, 45 m.

On all of the study sites thermokarst lakes and block erosion were observed; see Figure 10-4Figure 17, and Figure 19. The thermokarst lakes- or ponds were of varying size and some were covered in or surrounded by vegetation with distinct green vegetation. Also, they were very often located in direct contact with the palsas, or in the adjacent lee sides. Several investigations on measuring the depth of the ponds were performed. The majority of the ponds were > 1 m deep.

When squeezing a peat sample taken from the uppermost layers of a palsa top, and a wet mire sample, the volume decreased with more or less 70 %, meaning that air and moisture constituted the majority of the topmost layer. From this, a suggestion of the average palsa stratigraphy has been presented

From these stratigraphy and mire observations; three mire stratigraphy classes related to palsa mire was set. They were based on the apparent situation that there are large differences on soil wetness and stratigraphy governing within the mires in Finnmark. Hence, a wet mire stratigraphy, palsa stratigraphy and embryo palsa stratigraphy was set, and presented in Table 1.

The wet mire stratigraphy was interpreted to be consisting of mainly water in composition with vegetation from an old degraded palsa. The palsa stratigraphy was classified in two different scenarios, one for the general mature palsas and one for a typical embryo palsa, based on the field observations. The difference form the palsa and the embryo palsa stratigraphy are the height of the active layer and the height of the
ground ice. As the ALD in Lakselv was only 35 cm the ALD was set to 30 cm and thus somewhat downscaled. Hence, in the embryo palsa one is assuming the organic rich surface layer to be 20 cm lower than in the general palsa, and since the ground ice is still in the growing phase, the height of the ground ice was set to 20 cm. Assumptions are made below the active layer, and the ground ice layer (second layer in palsa- and embryo palsa stratigraphy) is assumed a height of 1.5 m and 0.2 m for the palsa and embryo palsa, respectively. This was based on the general height of the palsas and the active layer.

The assumption of a mineral layer with 5% volumetric organic and 45 % organic is also an assumption that this mineral layer is water rich and possibly in contact with the ground water table. A height of 14 meters for the mineral layer (layer three) is also an assumption; however one would assume this layer to be quite deep, and possibly also consisting of moraine as informed in 3.2. Nonetheless, the relative depth here did not change the model performance and the values have been comparable in similar studies (Brown and Romanovsky et al., 2008, Lawrence and Slater, 2008, Treat and Jones et al., 2016).
Table 1: Presenting the setup of Stratigraphy for the various modeled mire scenarios (palsa stratigraphy, wet mire stratigraphy and embryo palsa stratigraphy. Stratigraphies are based on lab results from the soil samples (Figure 12, Figure 13, Figure 14) and according to soil wetness interpretations from the field. For the topmost layer in stratigraphy 1 30 % is water, 5 % is mineral of silt, 15 % is organic content and 50 % is air due to the high porosity of the organic layer. While the wet mire stratigraphy is super saturated with water, no air is present.

<table>
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<th>Mineral Content</th>
<th>Organic Content</th>
<th>Soil Type</th>
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<td>0.05</td>
<td>0.15</td>
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</tr>
<tr>
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<td>Bedrock</td>
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Palsa

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</thead>
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<tr>
<td>1.0 – 15.0</td>
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<td>0.5</td>
<td>0.0</td>
<td>Silt</td>
</tr>
<tr>
<td>&gt; 15</td>
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<td>0.97</td>
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<td>Bedrock</td>
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</tbody>
</table>

Wet mire

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<td>0.05</td>
<td>0.15</td>
<td>Silt</td>
</tr>
<tr>
<td>0.3 – 2.0</td>
<td>0.5</td>
<td>0.45</td>
<td>0.05</td>
<td>Silt</td>
</tr>
<tr>
<td>&gt; 15.0</td>
<td>0.3</td>
<td>0.97</td>
<td>0.0</td>
<td>Bedrock</td>
</tr>
</tbody>
</table>

"Embryo Palsa"
Figure 15: Two field assistants measuring the ALD and geographic position and elevation of an embryo palsa in Lakselv. The embryo palsa was ≈ 50 cm, while the adjacent mature palsas in Lakselv were ≈0.5 m high. ALD was ≈ 35 cm. Notice the different colors of the vegetation on this palsa and the adjacent mire. Also the height of the vegetation is quite different on the palsa compared to the adjacent mire which consists of crowberries, cloudberries and blueberry bushes. The water body located on the upper right of the field assistants was interpreted to be a thermokarst lake. Photo from September, 2015 by Ingvild Solheim.
Figure 16: Embryo palsa in Karlebotn. Height was approximately 1 m and ALD was ≈ 45 cm. Surrounding palsas are several meters high >6m and have segregated ice below ≈55 cm. Photo from September, 2015 by Ingvild Solheim.

Figure 17: Thermokarst Lake and block erosion in Iskorás palsa mire. Notice the taller vegetation below the palsas, and in the surrounding mires e.g. birch tree to the left. Distance from thermokarst to palsa ≈ 20 – 30 cm. Vegetation on the palsa summit is dominated by a lighter grey color of lichens. Photo from September 2015, by Ingvild Solheim.
Figure 18: Thermokarst Lake in Iskorás next to a degrading and fracturing palsa. The thermokarst lake is in direct contact with the palsa. Photo from September 2015 by Ingvild Solheim.

Figure 19: Degraded palsa Karlebotn, presenting the typical vegetation on the palsas and the surrounding mire with smaller birch trees and shrubs on the edges of the palsa. Scale may be the active layer borrh in front of the distinct green vegetation in between the palsa.
5.1.3 Snow distribution and density

The winter fieldwork in mid-March 2016 was of great value for the model initialization and hypothetical reasoning for palsa distribution. Systematical snow depth measurements and snow sampling in coastal and continental palsa mires were performed. During this period the temperatures were relatively warm, barely dropping -7 °C with wind speed ranging from light to gentle breeze. During the last couple of days, however, March 15th -16th, a warm front arrived causing melting temperatures in Karasjok and Kautokeino (5 °C, with moderate breeze). This weather change was quite interesting and resulted in a severe change in the snow pack, which we were able to observe as we visited Aidejavri before the warm front and after. From Li and Pomeroy (1997) a threshold wind speed for snow transportation is to be 8.3 ms\(^{-1}\) for the mean ambient air temperatures for Kautokeino meteorological station, during the period of snow cover (02.11.2014-17.05.2015). The mean air temperature for the snow cover period was -7.4 °C. During winter 2014-2015 Aidejavri had a total of 46 days of wind speed exceeding this threshold.

From the in situ snow depth measurements substantial small-scale variations of snow distributions were observed on all of the study sites. For example, forested sites had 11 times more snow than palsa sites in Lakselv and in Soussjavri, see Figure 21. A large palsa in Aidejavri site is presented in Figure 20, together with the three sites where snow depth was measured. One may observe areas of bare ground on the palsa, in contrast to the forested areas and wet mire locations symbolized with yellow dots in the figure. Figure 23 and Figure 24, presenting overview of Lakselv study site, Oksebergmyra, on March 13\(^{th}\) 2016, also shows the trend of palsas with a much shallower snow depth than the lee sides and adjacent areas of the palsas.
Figure 20: Presenting a large-palsa in Aidejavri with the author as a scale (A) of 30×5 meter in length and height, respectively. The three yellow dots in photo A presents locations where snow depth- and density measurements (see Figure 21 and Figure 22) was taken. From left for right the yellow dots represent forest, wet mire, and palsa locations for the Aidejavri snow depth measurements. Photo B presents the low snow depth on the palsa (20 cm), and photo C presents the deep snow (60 cm) in the wet mire. Photo by Kristoffer Aalstad, 2016.
Figure 21 presents in situ snow depth and Senorge-snow depth from four of the five study sites during second week of March 2016. Snow depth on palsas varied from 0 – 20 cm, while wet mire snow depth varied between 55 – 80 cm and forest snow depth varied from 65 – 120 cm. Forested sites had substantial more snow, 4-15 times more, than palsa sites, and wet mire sites had 3 – 12 times more snow than palsa sites, and wet mire have about 15 – 33 % less snow than forest sites.

Soussjavri and Lakselv forested sites had the deepest snow, 120 and 110 cm respectively. While the palsa sites for the same mires were overlain by 10 cm snow or free of snow (Figure 21). It seems as though Aidejavri had the most evenly snow distribution for the three mire locations, as the difference between forest and wet mire snow depth were 15 cm and wet mire and palsa differed with 40 cm. After the warming period the snow pack evened out in the forest and wet mire, and the difference was even less (10 cm). The Soussjavri and Lakselv site however had the largest snow depth variation for the three mire locations. Also, the Senorge-snow was more than half a meter less than was observed in the forested sites.

All of the study sites show the highest snow depth in the forest, except Aidejavri after warming period. It seems as though the snow pack either compacted and -or melted 20 cm in the forest. The difference was less in wet mire site as it measured 5 cm less on the 15th of March. However the palsa snow depth had decreased with 10 cm in four days. Senorge-data only measured a five cm decrease.

Wet mire sites seem to be more comparative regarding snow depth, as it does not vary more than 25 cm from coast (Lakselv) to inland (Aidejavri). Soussjavri also has the deepest snow in wet mire sites, with 80 cm, while Aidejavri only measured 55 cm during the last measurement. Even though Lakselv palsas were bare blown, the wet mire sites had the second deepest snow with as much as 70 cm, which was 15 cm more than Senorge-snow measured at that time.

From this it is clear that the snow distribution in the four sites had similarities to one another. Snow depth varied greatly on a small-scale for all of the sites, as wet mire and palsa sites were located 1 - 5 meters apart and palsa and forest sites are 20-30 meters
apart. Senorge-snow depth varied from 50 - 75 cm, and the ratio between snow depth in forest and senorge-snow is greatest for Lakselv and Soussjavri (50-60 % more respectively). However Iskorás and Aidejavri (1) had the same ratio of 12 % less snow in the snow forecast from Senorge.no, while for the Aidejavri site, after period of melting temperatures, senorge-snow depth 5 cm deeper than the Forest site.

Figure 21: Snow depth measured on study sites. Measurements were taken on March 11th (Aidejavri (1)), 12th (Soussjavri), 13th (Lakselv), 14th (Iskorás) and 15th (Aidejavri (2)) 2016. Palsa in Lakselv were bare blown, thus a snow depth of 0 cm and no blue column for Lakselv site.
Figure 22: In situ snow density plots from Soussjavri (A), Aidejavri (B), Iskorás (C) and Lakselv (D). See legend for all sites on plot A. Aidejavri site have been measured before and after melting period, hence two columns from each site in this plot, see legend. Density is plotted in kgm$^{-3}$ for all sites. In Iskorás, snow density measurements were neither performed in wet mire nor on palsas, thereof only three columns representing snow depth in the forest.
From the snow depth sites, density measurements were also performed, presented in Figure 22. Unfortunately, Iskorás and Lakselv mires have not been measured on all of the three mire-locations, forest, wet mire and palsa. Snow density for Lakselv site has only been investigated in the forest, and Iskorás site have only been measured on palsa and- wet mire sites.

From the Soussjavri plot (Figure 22 -A) the forest site have the densest snow in the bottom layer with almost 275 kgm\(^{-3}\) while the bottom layer of wet mire site was measured to be 265 kgm\(^{-3}\). The top layer of Soussjavri forest site was approximately 255 kgm\(^{-3}\), wet mire topmost layer was 100 kgm\(^{-3}\) lighter, with 155 kgm\(^{-3}\) and the density of the palsa snow was slightly over 200 kgm\(^{-3}\). The mean density of Soussjavri sites was measured to be 245 kgm\(^{-3}\), 218 kgm\(^{-3}\) and 201 kgm\(^{-3}\) for the forest, wet mire and palsa sites respectively.

Aidejavri snow densities (Figure 22- B) have significantly different values before and after the days with melting temperatures, 195 -and 325 kgm\(^{-3}\) respectively for top layer of wet mire, 165 -and 250 kgm\(^{-3}\) for top layer in the forest and 145- and 325 kgm\(^{-3}\) for palsa. The wet mire site had very similar densities before the melting temperature period, 190-195 kgm\(^{-3}\), compared to 230-325 kgm\(^{-3}\) after the warming period. The largest difference before and after melting was measured on the palsa and the lowest difference was measured on the bottom of wet mire site. The lowest density in Aidejavri, of roughly 145 kgm\(^{-3}\), was measured on the palsa site, and the highest density, 325 kgm\(^{-3}\), was measured on the topmost layer of the wet mire after the melting period. The midst and bottom layer of forest site in Aidejavri were measured to be identical, 195 kgm\(^{-3}\). The same was also observed for wet mire, top and bottom layer, in Aidejavri with a value of 196 kgm\(^{-3}\) before melting period. The mean density was measured to be 177 kgm\(^{-3}\), 193 kgm\(^{-3}\) and 271 kgm\(^{-3}\), for forest, wet mire and palsa site respectively.

Iskorás densities have been measured in wet mire site and palsa site (Figure 22-C). The densest snow, almost 315 kgm\(^{-3}\), was measured on wet mire site at the topmost layer, while the palsa site was the driest site of almost 250 kgm\(^{-3}\). The density decrease with depth for the Iskorás site, with 315, 290, 255 kgm\(^{-3}\) for top, mid and bottom sample respectively.
Lakselv forest measurements show an increasing density with depth, with 310, 340, 350 kgm$^{-3}$ for top, middle and bottom sample respectively (Figure 23-D). The main density of the Lakselv forest site was 333 kgm$^{-3}$.

If we look at the study sites on a more regional scale with regard to snow cover, interesting differences was observed. Palsa sites in Lakselv mire were more or less free of snow, or constituted a thin layer of ice, see Figure 21, Figure 23, Figure 24. As this site is situated in the innermost part of Porsangerfjorden and more or less at sea level ($\approx 32$ m.a.s.l.), as well as surrounded by minor mountainous landscape with ridges in north and southeast ranging from 80-130 m.a.s.l., the site is susceptible to katabatic winds blowing down from the mountainsides, as well as temperate wind from the fjord. Wind speed in Lakselv during the last two decades have some gaps of missing data, however from Figure X it is clear that the mean wind speed is in the ranges of breeze (6 m/s). In Iskorás site palsas generally had a low layer of snow (Figure 23), while snow depth on wet mire sites had more than half a meter snow (Figure 24). The situation was somewhat different in the continental area of Finnmark in Aidejavri. From reference source not found., it is clear that the palsas were topped by 20 cm snow, however the snow depth were three- four times as much in wet mire- and forest sites. This study site is located in a relatively flat terrain varying from 360-430 m a.s.l., and mean wind speed for Kautokeino from 2007-2011 is approximately 3.5 ms$^{-1}$, which is according to Beaufort wind scale a light breeze. Nevertheless it is important to note the fact that these snow measurements were only performed during one specific period in one winter, and the situation may vary from year to year.
Figure 23: Overview towards west in Lakselv study site, Oksebergmyra on March 13th 2016. Notice the palsa (in the center of the photo) are more or less free of snow, while the surrounding mire and lee sides have a deeper snow cover. Photo by Ingvild Solheim, 2016.

Figure 24: Presenting the embryo palsa in Lakselv (Figure 15) during March 13th 2016. The palsa is more or less free of snow (where the field assistant is measuring the geographic position with GPS). Photo by Ingvild Solheim, 2016.
5.2 In situ snow distribution and active layer depth

Average Active Layer Depth

From active layer measurements performed summer 2015 a plot over the average active layer depth on palsas is presented, see Figure 25. The average ALD varied in a range from 0.48 – 0.60 m for the five study sites. The coastal mires, Karlebotn and Lakselv, and the continental mire Aidejavri had the deepest mean AL (≈55-60 cm). While the continental mires Iskorás and Soussjavri had the lowest mean AL (≈ 48-50 cm). Even though the study sites are plotted from the coast to more continental areas it is not possible to see a very clear trend in a decreasing of ALD with continentality. With a maximum and minimum interval of 12 cm, the average ALD in the studied mires are somewhat the same, however large local variations within each mire has been observed Figure 15Figure 16Figure 17Figure 18Figure 19.

Figure 25: Plot over empirically measured active layer in all study sites during September 2015. Notice the ALD does not vary significantly between the five sites.
5.2.1 Iskorás

The total number (n) of in situ points in Iskorás mire is 99, 44 of these points were located on wet mire, 4 were in the forest, which leaves 51 points on dry mire. Almost all of the in situ points are presented in Figure 26 and some are zoomed in on Figure 27 (A-F), and presented in histogram according to soil condition and snow depth, and soil condition and ALD (Figure 28). To get an overview of the snow and active layer depth in a comparable way, the points are classified in soil, wet or dry, and active layer depth (m), see legend. From this it is possible to extrapolate some local trends on the palsa mire.

Figure 26 (A-F) and Figure 27 presents some examples on what seems to be the general pattern for the Iskorás mire, and Figure 28 presents the same measurements in stacked histograms showing the relationship between wet and dry mire points and snow depth – and active layer depth (ALD). Dry mire points were measured to have snow depths ranging from 5- 58 cm snow depth, while wet mire points had snow depths of 1 meter or more. The dry mire points are mostly located on the palsas, while the majority of the wet mire-points are either situated on the outermost part of the palsa mire, in between palsas, or on palsa edges where the vegetation is light brown or green, Figure 26 (A-F).
Figure 26: In situ active layer –and snow depth measurements from Iskorás. Notice the different symbol and color for active layer depth and soil condition, see legend. The unit for the measurements are cm. Total number of in situ points in Iskorás mire is 99, however this Figure presents 94 points, as one point is situated further south in the mire and four points were situated in the forest where no permafrost exists and was thus excluded.
Sixteen points were measured to be of intermediate ALD (>55 < 100 cm) and situated in dry soil, which means 30% of the dry mire points were measured to have intermediate ALD. The majority of the intermediate points were also situated on the perimeter of palsas, especially in eastern and Southern direction, and a couple in the center in Figure 26, see zoom in on the eastern points in Figure 27. Some of these points, especially in eastern direction were also situated on central part of palsas (Figure 27). These intermediate points are somewhat located on the vicinity of palsa edge were the mire is still dry but denser than on the palsa top. One of the in situ points in Iskorás was classified as wet mire, however it had an ALD of 40 cm and was located on a narrow arm of a larger palsa, see zoom in on this point to the upper right in Figure 27. Points in wet mire with ALD >> 100 was situated in between palsas, and points in dry mire with ALD > 55 < 100 cm was often situated on palsa edges, while points with ALD < 55 were on palsas.

As Figure 26 shows, this mire is characterized by lakes or dams, measured to be up to 30 meters wide. These water bodies are surrounded by light brown and green vegetation which seems to be delimitating the transgression zone between palsa and dams. Two of the three dry points with AL >> 100 cm are located in this transgression zone, see upper right Figure 27.
Figure 27: Presenting a zoom in on the active layer measurements in Eastern part of Figure 26, with legend as in Figure 26.

One of the central transects crosses a 100 meter long palsa, where 9 points were measured, all of them were classified as dry mire. Seven out of the nine points had AL <= 55, and < 60 cm snow cover. The residual two points were either covered in deep snow or had an AL >> 100, and they were both situated on the perimeter of the palsa.

The six zoom-in inlets in Figure 26 also presents some interesting results regarding the positioning of dry and wet mire points, as well as snow depth variation. Some of the palsas in the mire were small and surrounded by water bodies and green – or yellowish vegetation. One example of this is presented in Figure 26-A. Here, one may observe the blue water body on the left side of the palsa, as well as the larger water body on the left in the figure. The two measurements taken in the yellowish vegetation in the same...
In Figure 26-B, one may observe a somewhat different situation, where the blue point in the middle had an ALD of 45 cm, however it was overlain by half a meter of snow and situated on the palsa perimeter. The two other points were either wet (left) or dry (right) with ALD >> 1m and covered with 55 cm and 30 cm snow, respectively. The right point was situated on the outermost edge of a palsa. The left point, which was in wet mire was located 20 meters from the right point. Indicating 25 cm snow difference, however the ALD was the same even though the points were situated in mire with very different wetness.

To sum up the in situ measurements from Iskorás, it is helpful to look at the two histograms in Figure 28. The snow measurements compared with active layer depth revealed that nine of eleven points with snow depth <=10 cm had an AL depth lower
than 55 cm, and was situated in dry mire. The residual points were both situated in wet mire had an active layer of 40 cm and > 100 cm. A total of 47 points was measured to have an AL >>100 cm. Figure 26 shows that 65 % (n=30) of them had a medium snow depth of > 10 < 60 cm, and they were mostly situated in the outskirts of palsas where the mire is wet. Only two of them were situated in dry mire. However 75 % of the points with snow > 10 < 60 cm had an AL of <= 55 cm. There were 23 points with deep snow (> 60 cm) cm and only five of them had an AL of < 100 cm, they were all situated in dry soil. Two of these four points were situated on top of palsas; one was situated in between two palsas (Figure 26-C) and two on palsa edge, see Figure 27. It seems that a snow depth of 80 cm is the maximum of snow depth for dry mire points in Iskorás, see Figure 28. Of the 44 points situated in wet mire, 32% (14) of them were overlain by deep snow, 64 % (28) of medium snow, and 4 % (2) of low snow. From the histogram plots it is clear that the wet mire points all have an ALD of >> 100 cm, except one single point which had an ALD = 40 cm and was covered in low snow. The majority of the points with high snow (> 60 cm) cover are all situated in wet mire, and the majority of the points with low snow cover are situated in dry mire.

The average snow depth for wet mire points with ALD >> 100 was ≈ 50 cm, while dry points with AL < 100 had an average snow cover of ≈ 30 cm. The snow depth seems to have a slightly decreasing trend with altitude and active layer in the mire, as the majority of the points with low snow depths and low active layers are situated on the highest altitudes, and the greater part of the wet points are located in the lower lying points in the mire where the snow height is somewhat deeper. The lowest point measured in Iskorás mire was however on 361,34 m a.s.l, and was situated on dry mire in the outermost part of the second transect. The ALD of this point was measured to be as low as 20 cm, and the snow depth was 65 cm.

From this it is possible to interpret that the coldest points in the Iskorás mire are located on dry mire, and warmest points are positioned on wet mire. And the snow depth increases with active layer depth, as the coldest points have lesser snow than the warmest points.
Figure 28: Stacked histogram plot presenting soil condition with snow depth (A) and soil condition with active layer depths (B) for the in situ points in Iskorás study site. The four points in the forest are excluded. Mark that the bins are separated in wet and dry soil, and the height of the bins sum up the total counts for one active layer or snow depth interval. None of the measured points had ALD in the interval 0-10 and 20-30 cm.
5.2.2 Soussjavri

The Soussjavri study site (Figure 29) is somewhat smaller concerning the total number of in situ points \( n = 37 \), however the two sites seem to have some similarities with regards to wet mire, ALD and snow depth pattern, see Twenty-four of the Soussjavri points were situated in dry mire, whereas the rest were in wet mire. The distribution is presented in Figure 29 and Figure 30. Also, the dry mire points seem to have lower snow depth than wet mire sites are mostly located on palsas. The opposite seems to be the situation for wet mire points, as they are mostly situated in between palsas or on palsa edges, however two points in transect 3 and one point in transect 4 seemed to be positioned on palsas. Also the easternmost point in the uppermost transect is located only 10 meters from high vegetation and forested area.

Four zoom-in inlets in Figure 29 presents some interesting situations regarding snow depth and ALD distribution in the palsa mire. Figure 29-A presents four points whereof two of them are situated on palsas with snow depth of almost 10 cm (left) and 0.5 m (right) and the other two on wet mire on palsa edges with deep snow of 80 cm (left) and 110 cm (right). The wet point on the right side in the figure is located on the palsa edge or in between two palsas in a depression. This Figure also demonstrates that the palsa is surrounded by water bodies on the long side to the left and also in the southernmost edge. Degrading of the palsa in the same locations are also possible to observe, as cracks or gaps on the palsa edge. This is also the situation in the central part of transect two (Figure 29-B), which presents part of a palsa being adjacent to vegetation with distinct green color, and a successive water body. Four points were measured in this area, three of them were situated in the green vegetation-area which was wet, and one was on the dry palsa.
Figure 29: Soussjavri Palsa mire with in situ measurements taken in 4 transects from left to right in the Figure. Four zoom-in figures are presented marked with letters (A-D). The total number of in situ measurements for Soussjavri mire is 37, and all of them are presented in the topmost figure. See Legend.
The four transects in the Soussjavri mire show a gradient of increasing snow depth from palsa to wet mire points. This is also possible to see in Figure 29-B, where the snow depth increases from 23 – 70 cm in a 40 meter distance from palsa point to the outermost red point. The palsa point had low snow cover (23 cm) and low ALD (50 cm), while the three wet points had different snow classifications (30 – 35 and 70 cm from left to right) and ALD >> 100. Thus, the two former wet points were situated in dry mire; however the snow depth was relatively shallow. One more point in Aidejavri was measured to be in the same situation, with a snow depth of 27 cm and ALD >> 100 cm. This point is presented in Figure 29-C, one may observe that it is situated on green vegetation in the middle of two palsa edges. The two other points are situated on a palsa (left) with dry mire, and near the palsa edge (right) on wet vegetation with a distinct green color. The dry point was underlying 30 cm snow, and was measured to have an ALD of 50 cm, whereas the wet point had more snow (50 cm) and ALD >> 100 cm.

As mentioned, the snow depth varied greatly on short distances from wet mire locations to palsas. However, the snow depth varied on top palsas as well. This is presented in Figure 29-D, where three dry points situated a few meters from the palsa edge is symbolized different snow depths. The point to the left was located closest to the palsa edge and was covered by 45 cm snow and had ALD = 75 cm, the other two had 35 and 25 cm snow, and ALD of 50 cm. This Figure also shows a palsa edge with fractures and degrading forms which borders to green vegetation and further north a water body.

From Figure 30-B it is clear that the wet mire points mainly has deep active layers >> 100 cm. However, two wet points were measured to have ALD of 90 cm and 20 cm. These two points were 55 meters apart with a height difference of 30 cm, located on the eastern part of transect 2 and 3 in Figure 29. They were covered with very different snow depths; the former point was covered with almost 110 cm snow (transect 2), while the latter point had almost 45 cm snow (transect 3). Meaning the point with the most snow also had the deepest AL, and was located on a lower elevation. One particular bar in Figure 30-B stands out, regarding dry mire points. This bar counts twelve points in the 40-50 cm bar. The 50-60 cm interval counts four dry points; they have snow depths ranging from 5- 43 cm.
The average snow depth for wet points with deep ALD (>> 100 cm) was ≈ 70 cm while dry points with AL < 100 had average snow cover of ≈ 35 cm. From the histogram plot (Figure 30-A) it is clear that points with relatively low snow depth (< 60 cm) are mainly classified as dry mire points, and vice versa for points with high snow depth. The histogram peaks on the third bar, which includes snow depth from 20-30 cm, where six of eight points are located in dry mire. However wet mire points does not seem to dominate any of the snow depth bars, but the six bars in the 60-120 cm interval is mostly governed by wet mire points.

The highest snow depth for dry mire points with ALD <= 55 cm was measured to be ≈ 80 cm, and the lowest snow depth measured in Soussjavri was 5 cm. The latter point was also the one with highest altitude (319.93 m .a.s.l.). The point which was lowest in the terrain was the first point in the 3rd transect from the left (316.98 m .a.s.l.). This point was also measured to have the highest snow depth of all the Soussjavri points (117 cm). The wet mire points are generally situated lower in the terrain than the dry mire points. No wet mire points are measured to be over 317.5 m.a.s.l, and the lowest dry point was located on 317,22 m.a.s.l with 27 cm snow. The snow depth seems to be slightly decreasing with altitude, as the snow depth on the wet points generally was deeper than on the wet mire points.
Figure 30: The 37 in situ points from Soussjavri mire is presented in two different stacked histograms, showing the statistical distribution between wet and dry mire and snow depth (A) and active layer depth (B). See legend. The bar interval on the x-axis is 10 cm, and the units below the tick marks represent the mean value of each bar. Plot B has an x-axis ranging from 20 - > cm because none of the points in the four Soussjavri transects had an ALD < 20 cm. None of the measured points in Soussjavri had snow depth in the interval >80 <90 cm.

5.2.3 Aidejavri

The Southernmost study site, Aidejavri (Figure 31) had some similar results as Iskorás and Soussjavri regarding ALD, soil condition and snow depth distribution. However this study site had a deeper snow cover than the other study sites, and the ALD seems to be higher for the dry points. The mean snow depth from the in situ points here was 64 cm. For the dry points alone it was 55 cm while for wet points the average snow depth was 72 cm. Half of the in situ points were situated in dry mire, and the other half was in wet mire.
Figure 31: Presenting the 27 in situ points from Aidejavri palsa mire with four zoom-in figures marked with letters (A-D). The points are distributed in three transects.
Figure 31-A presents three points which are almost 25 meters apart from each other. The two outermost points were both covered with 80 cm snow, while the snow depth on the middle point which was situated on the palsa was 30 cm lower. The point to the right was classified as wet mire with ALD $> 100$, and the vegetation here was light green—brownish. This point is further zoomed in on in Figure 32-A. Here it is clear that the point is located in between two palsas, 2-3 meters from the palsa edges. The other two points were situated on palsa with dry mire, which is of grey color in the orthophoto, and the ALD here was also measured to $> 100$ cm. The blue point in Figure 32-A was located on a palsa, had a snow depth of only 25 cm and an ALD of 45 cm. This point was measured to be situated 1 meter higher than the red point in the same figure. Hence, the snow depth varied with almost half a meter with 13 meter distance and 1 meter height difference.

Some of the same pattern may be seen in Figure 31-B where one point with deep snow (70 cm) was situated on a palsa, and had an ALD $> 100$ cm. About 30 meters to the North West from this point, two other in situ points was measured. They were located on almost the same altitude in dark green vegetation which surrounded a water body. One of the points had ALD of 65 cm, while the other measured $> 100$ cm. The former point was covered with over 90 cm snow, while the latter had 70 cm. Hence, this point had the same snow depth as the palsa top in the same figure even though the palsa was 1 meter higher than the latter. The two lower points in this figure has been further zoomed in on in Figure 32-B. The vegetation is more apparent in this figure. One may observe that the vegetation between the two points seems to be of heather and mash growths which have a greener color than the palsa edge. On the leftmost side of this figure, the vegetation has a light brown color which transgresses to even lighter and green color.

Figure 31-C presents four points which all have deep snow cover (65 – 75 – 78 – 65 cm, from left to right). The dry point was the one with the deepest snow, and had an ALD of 75 cm. The altitude of this point was also the highest, and the two lowest points were the two with the lowest snow depth. As the orthophoto in Figure 31-C shows, the three wet mire points are located either in a water body (left point) or in brownish – green vegetation in between palsas (middle and right point). Degrading palsa edges which borders to water bodies may also be observed in this figure to the left and bottom. The
left point is located right on the water body, almost half a meter from the degrading and fractured palsa edge. This point is further zoomed in on in Figure 32-C, here it is clearer that the point is located in a water body and one may see the cracks and degrading forms on the palsa edge which mounds out into the water.

An example of ALD-increasing with distance to palsa may be observed in Figure 31-D, where three points are zoomed in on. One point is located in dry mire on a palsa and ALD here was measured to be 45 cm. Next to this point one may observe a small darker circle (3 x 2.5 m) which seems to be fracturing. The two latter points are located in light brown wet vegetation, and both have ALD of >> 100. Thus, the snow depths for all three points were more or less 60 cm and the altitude difference between wet mire points and dry mire point was 1 meter. Between the two wet mire points one may observe a water body which transgresses into brown – greenish vegetation where the two outermost points were located. Taking a closer look at the palsa edge in this figure, one may observe some darker areas which may be fractures or degrading forms of the palsa.

As in the Soussjavri and Iskorás mire, Aidejavri mire also has several areas with distinct green vegetation which transgresses into water bodies in one direction and into dry mire or palsas in the other direction. Three wet in points such a transgression zone has been zoomed in on in Figure 32-D. Two of these points (left and right) were located right on the palsa edge, and the middle point was situated in the transgression between distinct green vegetation and light brown vegetation. The middle point was underlying 65 cm snow, which was up to 20 cm less than the two latter points. However, the altitude on the three points only differed by a few centimeters. Taking a look at the right point, one may see that the palsa edge her is directly in contact with water and several darker lines, indicating fractures are present. All three points in this zoom in measured an ALD >> 100 cm.
Figure 32: (A-D): Closer zoom of three of the four outlines in Figure 19. 20-D is outlined in Figure 19 and the legend is the same as in Figure 19. Notice the massive palsa fracturing in B, C and D.

From the statistical histogram plot in Figure 33-B it is clear that the wet mire points are outnumbering the dry points in the 95 to >> 100 cm ALD bar. None of the wet mire points had an ALD < 95 cm. However, four of the dry mire points had an ALD >> 100. They are zoomed in on in Figure 32-A and Figure 32 –B, one may observe that they are located in the transgression zone on the perimeter of the palsas. The majority of the dry mire points had an ALD of <= 65 cm. The lowest ALD measured in Aidejavri mire was 40 cm.

Regarding elevation, the dry points were the ones with the highest altitude. The highest point measured was 380.63 m a.s.l; this point was covered with 45 cm snow and had 45 cm ALD. In contrast to the lowest located on 378.98 m a.s.l, covered with 65 cm snow and had an ALD >> 100 cm. The 10 highest points were all dry, and it may seem as though no dry mire points were present below 379.95 m a.s.l. The snow depth had a vague degrading trend with altitude.

From the snow depth histogram (Figure 33-A) one may see that the dry mire points are actually evenly distributed in the range 10-80 cm, however no dry mire points are measured to have snow depth in the 45-55 cm interval. The wet mire points seem to increase with snow depth, but only one wet mire point is present in the 85-95 cm bar.
The majority of the in situ points had ALD >> 100 cm (Figure 33-B) whereof 14 points were located in wet mire. The dry mire points were measured to have ALDs within the ranges of 45 to 80 cm, and four dry points had ALD >> 100 cm. However the main trend is clear; wet mire point has ALD >> 100 cm, and dry << 100 cm.

Figure 33: The 27 in situ points from Aidejavri mire is presented in two different stacked histograms, showing the statistical distribution between wet/dry mire and snow depth (A) and active layer depth (B). See legend. The bar interval on the x-axis is 10 cm, and the units below the tick marks represent the mean value of each bar. The snow depth histogram (A) has an x-axis ranging from 15 cm because none of the points in the four transects had ALD < 20.
5.3 Model results

CryoGrid 2 simulations have been performed on different scales and regions. Section 5.3.1 to 5.3.3 presents runs with stratigraphies from Table 1 and snow forcing in conjunction with field observations as presented in section 5.1.3 and 5.2.1 to 5.2.3 (Figure 8 and Figure 9). From this, a palsa run, wet mire run and embryo palsa run has been performed.

The palsa and wet mire runs have been compared with ground temperature from the GST-loggers in Lindejaeggi mire near Iskorás. Two palsa runs are presented in section 5.3.1; both of them are initialized with air temperature from Iskorás borehole (A3) however the snow forcing is different. One run was forced with downscaled snow from Senorge, and a second run with palsa snow. The GST-temperature from a logger situated at wet mire is presented in 5.2.2, together with a wet mire run forced with air temperatures. In section 5.2.3 the embryo palsa run is presented. This run has been initialized with embryo palsa stratigraphy (Table 1), and a climate forcing from Lakselv Banak meteorological station, because it was at this site, Oksebergmyra, the author first discovered a possible embryo palsa in formation. Because, the senorge snow did not produce any MAGST <= 0, a palsa snow was used in this run.

5.3.1 Palsa run

Temperatures at the bottom of snow have been recorded at six points in Lindejaeggi mire. One of the loggers was placed on a palsa, and recorded GST during August 2014-August 2015. The temperature curve from this logger was plotted with temperature from a borehole station on the Iskorás Mountain, “Iskorás Air A3” logger (Figure 22), together with two palsa runs; palsa snow scenario, with snow depth and duration as in Figure 9 and Iskorás downscaled snow scenario, with snow depth and duration as in Figure 8. The measured GST and air temperature curves deviate somewhat as the air temperature drops down to -45 °C in mid-January, while the coldest temperature recorded in the GST was -13 °C, in late December. Snow depth on this site was measured to a maximum of 55 cm in beginning of January. The first snow fall was recorded on October 28th that season, with 5 cm, and the ground was bare again in late May
However, when simulating GST-temperature using air temperature forcing with the senorge snow depth, the temperature is too warm during the winter months, compared to the observed GST-temperature. Meaning the snow depth from Senorge was either too high or of too long duration, isolating the ground from the sinking cold air temperatures.

A run with downscaled snow was therefore performed, in order to give an assumption on how the snow depth may have been during the logged period. This run is also presented in Figure 22, and it is yielding a remarkably similar temperature curve as the observed GST-temperature. The snow depth in the downscaled run was more or less of the same height as the senorge snow, and the first snowfall was on the same date. However, the length of the snow season was shortened with two weeks, as all of the snow was set to be melted within the first week of May. While Senorge observations still measured over 40 cm snow in this period.

The resulting annual ground surface temperature from the downscaled run is surprisingly well fitting with the observed GST temperature, especially during mid-October – mid-January. In this period the curves are more or less overlapping one another, with temperatures well below freezing. As the snow depth increases in January and February, the temperature curve from the downscaled snow somewhat stabilizes (Figure 22). It has a less fluctuating temperature, even though the air temperature has large temperature gradients dropping from 0 °C to almost -30 °C within short periods. The observed GST curve on the other hand, slightly oscillates with the air temperature. Yet, as the air temperature has large temperature fluctuations, the GST curve does not drop under -13 °C from January onwards and does not exceed -2 °C, indicating that the uppermost part of the soil stays frozen. The minimum GST-temperature occur in late December, at -13 °C, when the air temperature was ranging down to -35 °C, which was one of the colder temperatures measured. From this, a temperature difference of ≈ 20 °C between air and ground surface temperature was measured.

Taking a look at the thawing season, the difference between modeled and measured ground temperature is greater than during the period of snow cover. As soon as the ground is free of snow in May, the modeled temperature suddenly starts to have great variations, and follows the air temperature logging. The situation is the same during late
summer and autumn, before the snow covers the ground. Also for the measured GST-temperature the curve follows the air temperature, however the oscillations are not as big, and the temperature does not drop under freezing before mid-October, as the air temperature does already in September. Nevertheless, because the temperature oscillations are so big, up to 40 °C temperature difference in a few days, and the measured GST temperature varies no more than 10 °C during winter (January) and about 10 °C (late June), the mean annual temperature for the modeled and logged ground surface temperature are 1.33 and 1.15 °C respectively. This is almost 1 °C warmer than the MAAT from the air temperature logger, which was -0.35 °C.

All though the curves from MAGST from the downscaled snow run and measured GST are in comparison with each other, one may notice that the modeled curve is somewhat warmer during mid-February – start of April. This may be indicating that the modeled run may represent a snow height that is too high during this time, in comparison to actual circumstances during winter 2015. Nevertheless, as snow depth measurements in Lindeajaeggi during winter 2015 was not performed; one can only assume this was the case.
Figure 34: Air temperature logger (A3), GST logger on a palsa and simulated GST temperatures, all from Iskorás site, with adjusted snow cover from period August 2014 – August 2015. Temperature curves are plotted with different colors. See legend.

During field work in Lindejaeggi in March 2016, a snow depth of 5 cm was measured on the exact location of the GST-logger. Remembering the snow depths measured in Iskorás mire (Figure 21, Figure 26) this was more than half a meter less than both wet mire sites and senorge snow. Therefore, a palsa run with typical palsa snow, obtained from the overall snow depth measurements in Figure 22, was also performed. The objective from this run was to better understand the temperature regime one can expect if the snow depth is lower than the senorge snow and even the downscaled snow.

From Figure 34 one may deduce that the palsa snow run was also within reasonable comparison with the logged GST. In fact, this temperature curve seems to be fitting the logged temperature even better than what the downscaled temperature curve does. Taking a look at the GST and palsa snow curves in May-October, the latter curve seems
to have a somewhat warmer temperature, and during some periods (mid-August to October) a larger temperature variation than the latter curve. The palsa snow curve too follows the air temperature, however it does not exceed 22 °C as the downscaled snow curve does (up to almost 30 °C), or drop under -20 °C. During the winter months when the ground is covered with 10 cm snow, mid-October to start of May, the temperature decreases to below freezing, and more or less holds negative temperature during the whole winter. Nonetheless, in the end of November the air temperature increases from -11.5 to 2.5 °C and the modeled temperature creeps up to 2 °C. In the coldest periods of the winter, end of December to February, the palsa snow curve is at its coldest and drops below -10 to -15°C. This is in contrast to both the GST and the downscaled snow curves, which are no colder than approximately -12 and 10 °C, respectively. Further into the spring months from mid-February to mid-April, the palsa snow curve and the GST curve almost overlap one another.

An AL depth of 0.55 m was modeled with the downscaled snow and palsa stratigraphy (Figure 23). This is coinciding very well with the in situ AL measurements from Iskorás. The TTOP was -0.27 °C, MAGST was 1.40 °C and MAGT equaled -0.23 °C. A thermal offset was modeled to -1.63 °C while the surface offset was -1.75 °C. However, from this one may reflect that the MAGST is not within the temperature range of permafrost definition of <= 0°C.

Figure 23 and Figure (Palsa run with palsa snow, MAGT month) indicates that a low snow run is yielding an AL of 0.51 m, which is somewhat shallower than with the downscaled snow depth. However the difference is not drastic. The TTOP was -1.34 °C, the MAGST was -0.06 °C, the MAGT equaled -1.65 °C. Thermal offset and surface offset was -1.59 °C and -0.30 °C, respectively. From this, and the palsa snow curve (Figure 22), one may infer that the temperature cools with lowering of snow depth, as the MAGST in the palsa snow run was actually below freezing and within the temperature definition of permafrost. The thermal offset and the surface offset however decreases in the palsa snow run.

A comparative statistical analysis was performed on the measured and modeled GST curves. The RMSE between measured and downscaled GST was calculated to 4.93 °C,
and the coefficient of determination was 0.52. While for the measured GST and palsa GST the RMSE and coefficient of determination was 3.72 °C and 0.76 °C respectively.

![Temperature graphs showing depth and temperature variation](image)

Figure 35: Palsa runs with two different snow scenarios. Upper two plots present MAGST temperature and mean annual temperature at thaw depth with palsa snow. The lower two plots presents MAGST and mean annual temperature at thaw depth with downscaled Senorge snow depth. The average temperature at the specified depth and a 0 °C line is also drawn, see legend.
Figure 36: Mean monthly ground temperature for selected months, from palsa run with Iskorás air temperature forcing and palsa snow. Notice the different ground temperatures in the active layer throughout the year. August temperatures are the warmest, while in October the temperature is close to freezing and in the coldest month, February, the temperature at the ground surface peaks ≈ -4 °C. At 5 meter depth the temperature profile is < 0 °C.

5.3.2 Wet mire run

A GST logger installed on a wet mire location in Lindejaeggi has been measuring the bottom temperature of snow during the period of August 17th 2014 to September 4th 2015 on a selected palsa and adjacent mire area. A plot from the measured GST temperatures in a wet mire point one year from August 17th shows that the temperature has great diurnal variations in the autumn and summer months, when the ground is free of snow (Figure 27). From Figure 22 (Figure 35) and Figure 27, one may compare the air temperature and the GST- temperature during snow season, and conclude that the temperature under the snow is at roughly freezing or a bit below freezing during November – middle of May. There are at least two minor periods where the GST
temperature reaches -1 °C - -2 °C during the winter. Examples are in the very end of December when the air temperature is almost at its coldest, with -35 °C (Figure 34). At this time the wet mire GST temperature drops down from roughly 0°C to -2°C (Figure 37). Also, in late March, during a time when the air temperature cools in a matter of a few days from -2 to nearly -30 °C, the GST temperature decreased with roughly -2 °C from the freezing point.

The wet mire run was forced with the temperature series from August 2014- August 2015 from the air temperature logger at Iskorás. A snow forcing of “wet mire snow” was used. This run does not produce any temperatures below -3 °C during winter months, and the TTOP equaled 2.01 °C. From the temperature curves in Figure 37 one may see an indicator of the spring, summer and autumn season when the surface is free of snow, as the MAGGST varies several degrees within short periods, just as the air temperature does. While during the months with snow cover, mid-October to mid-March, the temperature is more or less constant, around or mostly below the freezing point. Further down in the soil, the temperature stays well above freezing during the whole year, however during thawing season in early May the 0.40 m depth curve is almost tangent to the 0 °C line (Figure 24). This indicates that one does not produce permafrost in soil with wet mire stratigraphy and snow depth accordingly. A snow depth of 60 cm was in fact measured during fieldwork in March 2015, hence this is considered to be a realistic assumption for snow depth forcing for wet mire sites.

Modeled and measured GST temperatures for wet mire sites seems to have a good correspondence to each other, especially during winter with a thick snow cover. However in several periods the modeled GST curve is warmer than the measured one. But, during the two autumn months (mid-September – mid October) the modeled curve is actually more or less constant at 1.9 °C, while the measured curve varies with the air temperature curve. The coefficient of determination ($r^2$) was calculated to be 0,68 and a root mean square error (RMSE) of 3.7 °C was yielded when comparing the measured and modeled wet mire GST.
5.3.3 Embryo palsa run

Remembering from section 5.2.1 there was observed a possible new palsa in formation in Lakselv. Unfortunately, no soil samples were taken at the embryo palsa, however in assuming the soil stratigraphy is the same as in the palsas where soil samples were taken; one may propose a stratigraphy for a new palsa. This is presented in Table 1, and the only difference between new palsa and palsa is the active layer depth. We assume the active layer depth measured on the embryo palsa in Figure (photo of new palsa in Lakselv) to be representative for new palsas in Finnmark. From this, a model initialization with CryoGrid2 with stratigraphy according to new palsa (Table 1) and
with climate forcing from Lakselv, a temperature regime for a new palsa is presented (Figure 25). A test of snow wetness has also been performed for this site, assuming very dry snow has a density of 200 kg m\(^{-3}\), and “normal” snow density is 300 kg m\(^{-3}\).

Taking a look at Figure 21, Figure 23 and Figure 24, one may observe that palsas in Lakselv were mostly bare-blow or free of snow during mid-March 2016. A run with the snow depth from Lakselv meteorological station did not yield any permafrost. This might insinuate that as the maximum snow depth that this snow series had, 55 cm, might have been a too high initialization for the temperature forcing in Lakselv. Also, the senorge snow depth was not consistent with the observed snow depth in the mire. However, when lowering the snow depth to a run with palsa snow, the situation is quite different, Figure 38. Active layer measurements in Lakselv mire revealed that a new palsa in this mire had an AL depth of 0.35-0.40 m. This is within the modeled AL depth (Figure 38). One may observe that the upper two left plots are very similar. Nonetheless, the lowermost plot has colder temperatures during beginning of November and late December - early January. Studying the two plots on the right side, one may see that the temperature curve with the driest snow is more or less constant, while the densest snow run have some variations during the winter.
Figure 38: Temperature at different depth for embryo palsa run with Lakselv air temperature forcing and palsa snow of 200 kgm$^{-3}$ (top) and 300 kgm$^{-3}$ (bottom). Plots on left side presents MAGST and plots on the right side present temperature at thaw depth for each run. The different snow densities yield more than 15 cm difference in AL depth.

From Figure 38 one may observe that the temperature curve for the densest snow drops down to about -2 °C in late March, and circa -1 °C in May, while the dry snow curve more or less stays at -0.02 °C the whole year round. Also, the thaw depths differ with 16 cm, as the driest snow yields an ALD of 0.59 cm. From this one may insinuate that the new palsa in Lakselv may be reliant on dense low snow in order to be able to evolve and survive.

5.3.4 Synopsis of model results

Palsa runs have been performed for all of the study sites, for different snow scenarios and compared to in situ active layer depths. The in situ ALD may be compared with the modeled ALD (Figure 25 and Figure 39). However, for Karlebotn, Lakselv, Soussjavri and Aidejavri, the in situ ALD is somewhat deeper than what is yielded with the model (Figure 39). The runs with palsa stratigraphy and senorge snow depths are in most
cases within the ranges of the in situ measurements. However, when using senorge snow forcing for the Karlebotn and Lakselv palsa mire, no permafrost is produced. However, all of the continental sites produce MAGST <=0 with wet mire snow depths; this is only achieved when using a five-ten year period of initializing in a steady state with the climate forcing. When the model is run for a longer period, the soil temperatures are slowly increasing and rise above a MAGST > 0 °C. From field observations (Figure 21) and Figure 34 one may assume that palsa snow forcing is a reasonable snow depth for most palsa sites in Finnmark. The wet mire snow forcing proposed in Figure 9 seems to be reasonable and comparative with snow depth measurements (Figure 21) and in situ GST measurements (Figure 37).

Figure 39: Active layer depth for all study sites from different snow scenarios, see Legend. Notice that all of the sites, except Lakselv, produce an ALD higher with the senorge scenario, than with the palsa scenario, and Karlebotn and Lakselv does not produce permafrost temperatures with the snow depth from senorge scenario. The snow forcing is according to Figure 8 and Figure 9 (palsa snow scenario).
5.4 Regional modeling
As a validation for the CryoGrid 2, a regional ground temperature model for Finnmark County has been performed. The model has been run with air temperature and snow forcing from senorge.no for palsa and wet mire stratigraphy (Table 1). Snow forcing according to field observations (Figure 9) have also been performed. Thus, a run with palsa stratigraphy and palsa snow scenario is presented in section 5.4.2. From these runs, one may present the mean annual temperature at 1 m depth, as well as active layer depth for each run. These results are presented in Figure 40, Figure 41 Figure 42 and Figure 43. The grid size of each cell is 1km x 1km. Lastly, the results in section 5.4.2 are compared with the mapped current distribution of palsas in Finnmark in section (5.4.3), that was recently presented by Borge and Solheim et al., 2016.

5.4.1 Senorge Scenario
Palsa
A regional model run performed for whole Finnmark and parts of Troms County, with parameters for palsa and senorge climate forcing is presented in Figure 27. The resultant 1-m temperature ranges from -4.45 to 3.03 °C. However, regarding only Finnmark County, the temperature actually does not descend -2.0 °C. The lowest 1-m temperature modeled in Finnmark was in the inner parts of Finnmarksvidda, ranging down to -2 ° C. The grid cells with lower temperatures than this were located in Troms County, near Goulasjavri (-2.4 °C) and South-East. For example on Kirkestinden (1677 m a.s.l.) and Mattagaisi (1636 m a.s.l), which is an area near Goulasjavri where smaller east facing glaciers are prominent.

An area in Finnmarksvidda of 4.46*10^6 km² is delineated by the -2 °C isotherm in Figure Figure 40-A. The lowest temperature within this area is around -1.3 °C, symbolized with slightly darker blue color than the surrounding grid cells, while the highest temperature was just under -1 °C. This area encompasses Iskorás, Soussjavri and Aidejavri. However, all of them are on the upper limits of this temperature isoline. 1-m temperatures in these mires were in the ranges of -1.0 °C and -1.05°C for Iskorás and Soussjavri respectively, and Aidejavri is just within the isoline with -1.09 °C. As Figure 40 displays, the coastal and easternmost area of Finnmark has warmer ground temperatures, symbolized with
red and yellow. Karlebotn and Lakselv have a mean 1-m temperature of 1.35 and 1.60 °C, respectively.

Taking a look at the thaw depth in Figure 38-B one may observe a very clear delimiting line of permafrost. The >> 1 m active layer areas, symbolized with red colors, embody more than half of Finnmark area. This area includes the southwester areas, and thus both Lakselv and Karlebotn, meaning that with the climate forcing from senorge, more than 1 m of soil in these areas thaw during summer season. The situation is very different as we move further into the continental parts of the county. Aidejavri had the lowest thaw depth of 0.45 m, Iskorás were similar to Aidejavri with 0.47 m but Soussjavri thaws even deeper, 0.7 m. From this Lakselv and Karlebotn represents areas with ground temperatures well above freezing, while Aidejavri and Iskorás are well within the in situ ALD. Soussjavri was modeled to thaw deeper than observed in the field, but still within the permafrost limit.
Figure 40: Presenting mean annual ground temperatures at 1-depth (A) and thaw depth (B) for Northern Norway (Finnmark and Northern parts of Troms). Notice the considerable difference in ground temperature from the coastal areas to the west and the continental areas in southeast.
Wet mire

An investigation on how the soil parameters affect the ground temperatures has been performed on a regional scale for Northern Norway. The wet mire parameters presented in Table 1 and climate forcing from senorge.no have been used in modeling mean ground temperature at 1-m depth and thaw depth in a Finnmark (and Northern parts of Tromsø). The wet mire properties yielded a 1-m temperature range of 3.34 °C to -0.25°C (Figure 41). This is almost four times warmer than the palsa scenario with the same climate forcing (Figure 40). The coldest areas, delineated with the 0 °C isotherm, were only produced in one grid cell in Karasjok (UTM33 907755.059, 773957.303) where the elevation is 125 m and (temp -0.001°C) and a few grid cells in mountainous areas of Troms with elevation > 920 m a.s.l. The coldest grid cell was modeled to be in a north facing river canyon, Reisadalen in Troms (UTM 33 769117.239 7708513.824) at ~1300 m a.s.l.

A belt stretching from west to east in the central parts of Finnmarksvidda is delineated by the 1 °C isotherm, see Figure 28. The elevation in this area is ranging from > 100 < 700 m, while the ground temperature ranges from 1 – 0 °C. Interestingly, only one of the study sites was within this belt. The three continental study sites were modeled to have a mean 1-m depth temperature of > 0.5 °C < 1.5°C. Iskorás were modeled to about 1.1°C while Soussjavri was again on the border of the 1°C, with mean temperature just under 0.95 °C. Further south, Aidejavri had a mean 1-m temperature of 1.2 °C.

More than half of the modeled area is within the 2 °C isoline, symbolized with yellow and orange in Figure 28. And the coastal areas are mostly within the 2- 3 °C range. This means that Lakselv and Karlebotn are again both within the warmest 1-m temperatures, with almost 2.2 and 2.5 °C respectively.

The wet mire parameters together with Senorge.no climate forcing does not produce ground temperatures that coincide with permafrost temperatures. Hence, no active layer plot is presented for this run.
Figure 41: Presenting mean annual ground temperatures at 1-m depth (A) and ALD (B) for a regional run for Northern Norway with senorge climate forcing and wet mire stratigraphy. Temperature is °C, and contour lines are set at 0, 1, 2 and 3 °C, see legend. Notice that none of the 1km² grid cells had mean 1-m ground temperatures below 0 °C; hence no thaw depth plot is presented for this run.

5.4.2 Palsa Snow Scenario

Palsa
As presented in section 5.3.1 palsa runs with palsa snow scenario yields colder ground temperatures than palsa runs with senorge snow or wet mire snow scenario. The modeled active layer for the study sites varied between 0.49 – 0.55 m (Figure 25), while the in situ measurements varied between 0.48 – 0.6 m (Figure 33 and Error! Reference source not found.). Hence, the modeled thaw depth only varies with approximately 5 cm for the 5 study sites. When CryoGrid 2 was forced with palsa stratigraphy and palsa snow scenario (Figure 42) the ground temperatures for Finnmark was within the ranges of 0 to -5°C. The average temperature 1-m plot (Figure 42-A) shows that ground
temperatures for palsas are in the range of (-3 – -5 °C) in the innermost parts of Finnmark.

When studying Figure 42, one finds that an isotherm of -4 °C is representative for the majority of the southeastern part of Finnmark (Finnmarksvidda). Hence, Iskorás, Soussjavri and Aidejavri study sites are all within the 1-m ground temperature of -3 to -4 °C. Interestingly, one of the coldest areas within the 4 °C isotherm occurred around Iskorás; which had a 1-m temperature of -4.0°C on the summit, and -3.5 °C in the palsa mire. Soussjavri site has a temperature around -3.7 °C and Aidejavri 3.1 °C in this run.

Some areas in high altitude areas were delineated by the -5 °C isotherm. These areas are located in Carravarri (800 m.a.s.l.), near the Southernmost tip of Finnmarksvidda (480-600 m.a.s.l), in mountainous areas of Troms County near Moskodal (775 – 1440 m.a.s.l), in the highest elevations in the National Park Stabbursdalen south west of Lakselv (850-1000 m.a.s.l) and some parts of Gaissane mountain range which represent the highest summits in Finnmark. Lower lying areas (0-300 m.a.s.l) stretching from south to east near the Finnish border was delimited by 1-m temperatures within the range of -1 – -3 °C. Consequently, Karlebotn study site had a 1-m temperature of -1.1 °C. The coastal areas are represented by warmer colors, and are within the 0.2 – -1 °C isotherm. The grid cell covering Lakselv palsa mire had a value of -0.2 °C.

Taking a look at Figure 42-B (thaw depth palsa low snow), interestingly enough the thaw depth ranges within 0.4 – 0.5 m for all of the continental study sites (Iskorás, Soussjavri and Aidejavri) as well as Karlebotn (0.47 m), while Lakselv is somewhat warmer as it thaws down to 1.2 m. The lowest thaw depth is modeled in all of the mountainous areas, and the deepest was modeled in the lower areas on the coastline.
Figure 42: Presenting the mean annual ground temperature at 1-m depth (top) and thaw depth (bottom) for palsa snow scenario run. The coldest temperatures are produced in the mountainous areas in southwest and in Gaissane Mountains west of Lakselv, and in a few grid cells near Aidejavri. The ALD is thus lowest in the high-elevation areas in southwest and within Gaissane Mountains (up to 0.4 m) while for the majority of eastern Finnmark this run produces an ALD of 0.4 - 0.5 m.
5.4.3 Palsa snow scenario compared with mapped palsa distribution

A map of the distribution of palsas in Finnmark has been produced by interpretation from aerial photographs (Borge and Westermann et al., 2016). Figure 43 presents the mapped palsas in comparison with the 1-m average ground temperatures. From this one may see that the palsas are mainly situated in a zone of 1-m temperatures within the isotherm delimiting -4 to 3 °C mainly. The majority of the mapped palsas (Borge and Solheim et al., 2016) are located within the southernmost part of Finnmarksvidda where the modeled palsa scenario also yields the coldest mean ground temperatures. A few grid cells around Karlebotn and Lakselv are also mapped to be consisting of palsa mires. However, as stated, these two latter palsa mires are within the warmest 1-m temperature isoline.

Figure 43: Presenting the MAGT at 1-m depth for the palsa snow scenario in relation to the recently mapped palsa distribution in Finnmark (Borge and Westermann et al., 2016). The percentage shows the concentration of 250 m² grid cells that have presence of palsas in 1km² grid cells. The concentration is highest in southeast of Finnmarksvidda, where also the lowest 1-m ground temperatures were produced.
6 Discussion

This chapter contains discussion and implications from fieldwork observations (6.1), which leads to a conceptual model for palsa degradation (6.2), discussion of the model results and uncertainties of the model performance (6.3) and lastly suggestions for further research (6.4).

6.1 Implications from field work observations

6.1.1 Properties of palsa mires

Taking a look at Figure 14 one may see that the volumetric organic content has a mean value (≈20 %) of a bit higher than what has been used in the model initialization (15 %). Increasing the volumetric organic content to from 15 to 20 %, the ALD did not lead to sufficient changes in the permafrost temperature for the stratigraphy forcing to be changed to 20 %. The very high DBD in Karlebotn topmost sample was interpreted to be of very compacted peat, and such high values are usually found much deeper in palsas (Oksanen and Kuhry et al., 2003, Kuhry, 2008). This may be due to a high compaction and might be indicating the Karlebotn palsas to be of very old peat compared to Lakselv site (Harris and Nyrose, 1992). Earlier studies have found DBD values for peat samples in same soil depth within the ranges of 0.07-0.27 g cm\(^{-3}\) (Kuhry, 2008) and 0.061-0.157 g cm\(^{-3}\) (Zoltai, 1991) in Canada and 0.12 – 0.26 g cm\(^{-3}\) in Finland (Oksanen, 2006). Hence, the Karlebotn top sample is somewhat skewed in relation to the other three peat samples and the latter studies. And due to the relatively good resemblance between modeled and measured active layer depth in Finnmark (Figure 25, Figure 26, Figure 31, Figure 35, Figure 37, and Error! Reference source not found.), one may insinuate that the soil forcing from both wet mire, palsa and embryo palsa is within consistence of reality.

From the total volumetric content plot one may see that the mineral content is very low, ≈5 %. These results are also in compliance with other studies concerning stratigraphical properties of peat, e.g. Kokfelt and Reuss et al. (2010) and from parameterization based on global estimates of soul carbon content (Lawrence and Slater et al., 2008b); they are much less than some as well (Westermann and Schuler et al., 2013). The latter study
relies on surface sediment classes by the Norwegian Geological Survey (Thoresen, 1990, Engevik, 2010), which was obtained from air photo interpretation and high-resolution mapping. The 50 cm high organic layer is comparable with palsa mires in Canada (Jean and Payette, 2014), however it was based on the observed peat thickness (ALD) in Finnmark. Thus, the in situ measurements may be interpreted to be more realistic. However, only four peat samples make out the basis for the soil forcing, this may be a shortcoming of this thesis and a source of error for the model results. But, the samples are from different mires and in different parts of the county, and the results are convincingly similar to one another. Hence, one may discuss that palsas have more or less the same stratigraphy, at least in the active layer.

Large differences between the soil samples were observed, concerning the organic matter content and the difference between the sediment sequence and the peat sequences. The organic matter content (LOI) for the four peat samples was very high and ranging from 98-97 %, while the sediment sequence from the river plateau had only 38 % LOI (Figure 12). The same methods as is described in section 4.1 (burning of soil samples) was also performed on core samples, from 0 - 355 cm depth, from peatland in the Abisko area in Northern Sweden (Kokfelt and Reuss et al., 2010) and in Canada (Kuhry P., 2008). Kokfelt and Reuss et al. (2010) found that a peat sequence had organic matter content (LOI) of 100 to 20 % from the palsa surface to about 100 cm depth, respectively. Similar values have also been found by Oksanen P.O. (2006) and (Kuhry, 2008), where peat samples had LOI of 89.8-99 % in 0-60 cm depth, which are in correspondence to sample depths in this thesis.

The latter authors and Lawrence and Slater et al., (2008) also found that LOI decreases with depth. From this, one may insinuate that the organic content decreases with depth. This was also the case for Lakselv samples, which had 1 % less organics in the bottom sample compared to the topmost sample. Conversely, Karlebotn actually had 0.4 % more organic in the bottom sample. Nonetheless, the Karlebotn samples were taken on shallower depths than Lakselv, supporting the former insinuation of a decreasing of organic with depth. The assumption that below 2 meters the soil consists of 5 % organics and almost 45 % minerals, and the rest is water in the soil forcing for CryoGrid 2 (Table 3) is in consistence with earlier deep cores from peatland in Canada (Kokfelt and Reuss et al. 2010). However, these cores of Kokfelt and Reuss (2010) are only 3.55
m deep, and the assumptions that bedrock is reached at >15 m depth should be checked with GPR. Nonetheless, such assumptions might not have a great meaning for the model results, as it is the importance of the peat cover over the mineral layer that yields the possibility of palsa development (Luoto and Seppälä, 2002).

The organic rich soil in Abisko area was found to be consisting of macro scale plant fossil and silt (Kokfelt and Reuss et al., 2010). This was also the case for the samples in Lakselv and Karlebotn, as one may see in Figure 10. The new palsa in Lakselv (Figure 15) was dominated by a carpet of S. lindbergii, this was also the case for palsas and peatlands in Hudson Bay area in Canada (Kuhry, 2008) and in Abisko area (Kokfelt and Reuss al. 2010). According to Sonesson (1970b) and Oksanen and Kuhry et al. (2003) vegetation carpets of S. lindbergii frequently occur around palsas, and may be used as possible indicators of permafrost in the near vicinity. This is supporting the theory that the small peat mound in Lakselv was actually permafrost, and embryo palsa in formation. But Kokfelt et al. (2010) observed this vegetation in a recently thawed palsa. From this one may alter the interpretation that the formation in Figure 15 and Figure 16 may not be an embryo palsa, but in fact a degraded palsa. However, due to the fact that a layer of ice was reached 35 cm down in the peat during boring on the embryo palsa in Lakselv, and neither water bodies nor fractures and depressions was observed on the two formations, one may still propose the peat formation to be an embryo palsa. Kokfelt and Reuss et al. (2010) revealed that a degraded palsa consisted of detritus gyttja from 0-320 cm depth and organic rich, silty clay from 320-355 depth. Detritus gyttja is organic rich sediment, formed in lakes, as it is a coprogenic formation with remains of dead organisms as plankton and fish. This may also be the case for both Lakselv and Karlebotn peat, as they are both located on the coast, and the fact that their core was consisting of silty clay supports the assumption of silt under the ground ice (Table 3).

The age of the palsas in Finnmark have not been studied in this case, however this would have been an interesting study and might lead to a better understanding of the paleoclimate in Finnmark during Holocene, as palsas and the succeeding vegetation which make up the peat are indicators of climatic conditions due to its preference for cool and moist conditions with a stable water table in order to establish (Zoltai and Vit, 1990, Zoltai, 1993, Oksanen and Kuhry et al., 2003, Oksanen, 2006, Kuhry, 2008). Estimation on the age of peat by their weight has in fact been done on peatlands in
Canada, by Zoltai and Vit (1990) and by $^{14}$C-dating by (Zoltai, 1993, Van Vliet-Lanoe Seppala, 2002). Zoltai (1993) found that permafrost degradation is initiated by sudden appearance of peat in the core sample which was deposited in very wet conditions. Zoltai (1993) also concluded that such degradation was initiated with fire, as one would observe traces from as layers of charred surfaced on a previously dry peat plateau. This is interesting, as such charred layers were also found in Lakselv (see Figure 10). Hence one may deduce that Lakselv mire have also had some periods of warmer climate, and possibly previous stages of palsa degradation. A degradation-aggradation period may be as short as 600 years, according to Zoltai (1993.). Nonetheless, one would suggest that the age of the peat in Finnmark is increasing with depth, as would be in consistence with literature. And the vegetation in such environments is of great diversity, as one may see in Figure 15,Figure 18 and Figure 19. The observed heterogeneity of the vegetation on the palsa mires in Finnmark are in consistence with several authors (Matthews and Dahl et al., 1997, Blyakharchuk Sulerzhitsky, 1999, Luoto Seppälä, 2003, Zuidhoff Kolstrup, 2005, Oksanen, 2006, Kokfelt and Rosén et al., 2009, Kokfelt and Reuss et al., 2010, Jean Payette, 2014). From peat samples in Sweden, Kokfelt and Reuss et al. (2010) presents a dating of a degraded palsa to 800-200 Cal yr. BP, and a palsa aggradation at 120 yr. BP. while in Canada Kuhry (2008) found that one peat plateau -and a palsa mire were in the ranges of 6585-470 BP.

6.1.2 Local factors controlling palsa existence

As the presented field work observations and measurements was conducted during one summer season and a few days during winter time, and only in selected areas of each mire with relatively low in situ points, the outcome of the snow forcing and interpretations from the in situ measurements lead to some uncertainties. In order to investigate the palsa evolution, how the snow distributes in the terrain from winter to winter and how the active layer evolves with the present climate, the measurements should be performed in successive years over a long period on the exact same location. This would yield a sufficient dataset of the representative snow scenarios in palsa mires, and the state of the ALD evolution. Just as has been done in Sweden (Åkerman and Johanson, 2008, Zuidhoff and Kolstrup, 2005), Finland (Seppala, 1982, 1990, 1994, 1997, 2002, 2003, 2008, 2008 and 2011) and Southern Norway (Solliid and Sørbel, 1998,
Gisnås and Westermann et al., 2014), Quebec (Vallee and Payette, 2007) and Alaska, Russia, Canada and central Asia (Lemke and Ren et al. 2007). A recent investigation on the development and distribution of palsas in Finnmark revealed a lateral degradation of palsas in the range of 33-71 %, in selected mires in Finnmark from the 1950’s to 2010’s (Borge and Westermann et al., 2016). Nonetheless, this work did not include systematic active layer measurements. But when combining this substantial degradation with results from this thesis one may infer possible controlling factors for palsa distribution in Finnmark, and discuss how it affects the state of permafrost in Finnmark.

If these mapped trends on thawing palsas continue, the measured active layer depth will increase, and permafrost temperatures as well. Such trends are already observed in e.g. Alaska (Osterkamp and Zhang et al, 1994, Romanovsky and Osterkamp, 1995, Osterkamp and Romanovsky et al., 1999), Europe (Frauenfeld and Zhang et al., 2004, Åkermann and Johansson, 2008, Lemke and Ren et al., 2007, Gądek Leszkiewicz, 2012) and in the Qinghai-Tibet plateau (Wu Zhang, 2008, Wu and Zhang et al., 2012) and are expected to continue in the future decades (Pavlov, 1994, Anisimov and Shiklomanov et al., 1997, Anisimov Reneva, 2006, Christiansen and Etzelmüller et al., 2010, Romanovsky and Smith et al., 2010, Vieira and Bockheim et al., 2010, Gisnås and Etzelmüller et al., 2013) and possibly even accelerate in the near future (Stendel and Christensen, 2002)

The Southern boundaries of permafrost have already moved northward in Canada by as much as 120 km (Kwong and Gan, 1994) and the ¼ of the Earth’s land area underlain by permafrost is expected to contract substantially in response to the climate warming (Anisimov and Nelson, 1996), however the rate at which it will decrease is ambiguous according to Burn and Nelson (2006). However, this is crucial news as the strongest climate change is expected Polar Regions.

Even though this thesis reports two findings of possible palsas in formation and the ALD measured on the embryo palsa in Lakselv is in consistence with other measurements on young palsas in Finland (Seppälä and Kujala, 2009), the conclusions from the summer fieldwork are nonetheless somewhat depressing. Large thermokarst ponds (Figure 10-D, Figure 17, Figure 18, Figure 19, Figure 26(A,B,D,F), Figure 27, Figure 29(A-B), Figure 31(B-D) and Figure 32(B-D) fracturing palsas (e.g. Figure 18 and 19, Figure 29(A,B,D) and abrasion of the peat was observed in all of the palsa mires (10-4). And as they are representing transects from west to east and north to south in Finnmark, one may
therefore imply the reported degradation to be the current situation for the palsas in Northern Norway, as would be in consistence with recent research in the area (Farbrot and Isaksen et al., 2013, Borge, 2015, Borge and Westermann et al., 2016).

Palsas are located within the southernmost margins of discontinuous permafrost zone (Luoto and Fronzek et al., 2004c), hence they are exceptionally sensitive to climate warming. When palsas reach the degradation phase with surface abrasion, fracturing and thawing of the internal ground ice, the exposition to summer temperatures and winter snow cover increases. Example of such fracturing and possible exposition to snow aggradation during winter may be seen in Figure 10-D. Such a situation may be leading to an escalated degradation of the palsas (Payette and Delwaide et al., 2004).

Thus the ongoing palsa and peat plateau degradation reported from the last six decades in Finnmark (Borge and Westermann et al., 2016) and in Canada (Payette and Delwaide, 2004) might escalate in the future as the mires gets warmer and wetter ice ground ice thaws with the projected continuity on climate warming. Excessive change on permafrost have already been reported for the areas covering the largest sources for organic carbon, and the consequences it has on hydrology (Luoto Seppälä, 2003) and vegetation (Christensen and Johansson et al., 2004, Bosiö and Johansson et al., 2012) not to mention anthropological hazards for infrastructure by subsidence by increasing of the active layer (Couture and Smith et al., 2003) and environmental effects of CH₄ and CO₂ emissions (Cao and Gregson et al., 1998, Repo and Susiluoto et al., 2009) are still not fully understood and might have been underestimated in earlier work. In Canada, roughly 50 % of the ground surface is within permafrost regions (Couture and Smith et al., 2003), and the infrastructure in northern Canada relies on stable temperatures in the frozen ground in order to prevail in the landscape. The disappearance of permafrost in these inhabited areas will have costly outcomes, as anthropogenic constellations as pipelines and roads are elemental in our everyday lives.

According to (Washburn, 1980) palsas are generally restricted to areas with MAAT <= 0 °C. Both Lakselv and Karlebotn were in areas of almost 1 °C during season 2014 – 2015 (Figure 6), and the MAAT from the two periods 1966-1983 and 1998-2014 (Borge, 2015 p. 44) show that Lakselv have had a warming of almost 1°C from the former to the latter time interval, indicating these palsas are likely in a degrading phase and that the
mapped degradation will continue. The continental mires on the other hand, are within Washburn’s (1980) 0 °C isoline, however the MAAT have increased from -2.59 °C to -1.51 °C during the last 47 years (Borge, 2015, p. 46). Luoto and Fronzek et al. (2004c) found by a spatial analysis that in Fennoscandia that palsas are located in a temperature window of -5 °C to -3 °C and precipitation below 450 mm yr⁻¹. In Finnmark, palsas are mainly located in areas with MAAT -4 – -3 °C (Borge and Westermann et al., 2016). But, if the palisa snow scenario as measured in the field continues to be the current snow depth on palsas in these mires, the palsas may actually be in a steady state in the future, just as have been modeled with CryoGrid 2 in section 5.4.2. Nonetheless, due to the massive lateral erosion and thermokarst lakes observed here, the state of art for the mires might as well be of a degrading phase.

Taking the illustration by Seppälä (2006) into account (Figure 2); palsas collapse as they are old and too high, and thus undergoes lateral erosion leading to exposing the internal peat and ground ice, resulting in a complete thaw of the ground ice and in the end to a thermokarst pond. Thus the theory of a cyclic evolution of palsas (Figure 2, by Seppälä and Kujala (2009)) may be the case for the palisa mires in Lakselv and Karlebotn. This would indicate that permafrost is currently forming with the present climate at these study sites.

From the in situ measurements of snow depth, snow distribution and active layer depth, there seems to be strong indications towards clear local trends in palisa mires in Finnmark, northern Norway (Figure 10-14 and 17-21). Firstly, soil dryness and palsas seems to be correlated to each other, as none of the dry points were located elsewhere than on the palsas. And the wet mire points were more or less located right next to palsas -or in between palsas (Figure 29-A-D). Henceforth, one may assume that dry peat is more or less restricted to the palasa summits. This may be explained by the height difference between palasa summit and wet mire sites. As the frozen core of peat leads to buoyancy of water percolating downwards in the summer when the active layer thaws, and further uplifting of the palasa summit as the water freezes. This uplifting of the palasa leads to an increase of the exposition to wind which dries the upper peat layer during summer. This was also found in palsas in Finland (Seppälä, 1982, 1994, 2003), and the process is illustrated in Figure 2. Precipitation in the form of rain will either be dried out by wind (Seppälä, 2002), or drained towards the bottom of the ground ice in the palasa,
or towards the lower lying wetter areas. However, the classification of humidity has some possible errors due to the fact that the observations were interpreted subjectively in the field, and not in the lab. And the classification is based on samples from the subsurface, and may be skewed by recent precipitation. However, since these were the large-scale trends on all of the sites, the assumptions made in the field regarding soil wetness seem reasonable.

Moreover, both the vegetation and snow cover between dry mire sites and wet mire sites are significantly different. Such tendencies have also been found by other authors in Northern Sweden (Sonesson, 1970a, Zuidhoff and Kolstrup, 2005). By snow depth measurements across palsas, the latter authors revealed that the palsas were almost free of snow in Laivadalen, and Seitajure. Laivadalen has almost two times more winter precipitation compared to Seitajure (343mm compared to 188mm), but is more or less without snow, while the Seitajure palsa were overlain by snow depth in the ranges of 5 to 60 cm. On the surrounding areas of the two palsas, snow depths were almost 50 cm in Laivadalen and up to 80 cm in Seitajure. Remembering the figures in section 5.2 one may infer that observations in all of the study sites for this thesis are in coincidence with that of Zuidhoff and Kolstrup (2005). For example palsas in Lakselv were in fact free of snow (Figure 21, Figure 23, and Figure 24), while lee sides measured 70 cm snow (Figure 21).

Aidejavri too has a very clear snow distribution, as there was not measured one single wet point with “palsa snow” (<10 cm) and all except one point had “wet mire snow” (>60 cm). These systematic snow depth measurements shows that there are major differences in snow cover between pal sa summits and the surrounding areas, and that the summits may well be without snow even though the winter precipitation is of considerable amount Figure 8. This was also reported to be a standard scenario in palsa mires by Seppälä (1990, 1994).

This means that other factors than the amount of precipitation is controlling the snow cover in the study sites for this thesis, such as wind drift of snow, as has been reported in earlier work in Scandinavia (Seppälä, 1982, 1986, 1988, Vliet-Lanoe and Seppälä, 2002). In the late 60-70s Seppälä (1982) experimentally demonstrated that by removing snow from a palsa several times during three consecutive years, a perennial frost layer was formed, already after the first winter. This may explain the abundance of palsas in
Karlebotn and Lakselv, which is located in areas with MAAT and with sufficient wind to keep the palsa more or less without snow during the winter. He also found deeper snow in thermokarst lakes in the adjacent areas of the palsas, than on the palsa summit. From Figure 41 it is clear that a typical wet mire snow depth does not produce perennially frozen soil, and the temperature is around freezing during snow cover months. This has also been found in Finland, where a 1-m thick snow cover was such a good insulator for the GST that the temperature remained at a constant 0 °C during the period of snow cover (Seppälä, 1990).

The local factors controlling palsa formation in the study sites in Finnmark seem to be somewhat different from one another. Lakselv and Karlebotn are located merely at sea level, only a few hundred meters away from the ocean, and have a higher MAAT than the continental sites. Unfortunately, winter fieldwork has not been performed in Karlebotn; however during the last three winters the maximum snow depth was measured to be over 60 cm according to senorge data, and the snow melts around the same time in all three years (late March to first week of May) (Engeset and Tveito et al., 2004a, Mohr Tveito, 2008, Saloranta, 2012), indicating a snow depth similar to the other sites, see Figure 8.

Wind direction is almost the same for the two coastal sites, from South. However in Karlebotn almost 50% of the wind speed is in the range of 5.3-10.2 ms⁻¹ (Figure 49-A), while for the southern winds in Lakselv, 20% are measured within the ranges of 5.2-10 ms⁻¹ where 10% is 7.6-10 ms⁻¹ (Figure 49-B). These wind velocities are according to Li and Pomeroy (1997) enough to transport and redistribute the snow in the windward direction (S-SW), leaving palsa summits more or less free of snow and the lee side of the palsas with typical “wet mire” snow. Though no systematic mapping of snow depth on lee sides were performed during field work for this thesis, one would suggest that the snow depth to be deepest on north facing lee sides of the palsas by the prevailing wind directions. From section 2.2.4 and earlier work (e.g. Seppälä 1982, 2002) it is clear that wind is a clear parameter for palsa formation. One may therefore insinuate the coastal palsa mires in Lakselv and Karlebotn to be dependent on sufficient wind speeds in order to survive the present climate. Goodrich (1982) actually found that the MAGT increased with several degrees in the presence of snow cover. And an increase of winter snow cover in Russia has led to a substantial increasing of the thaw depth in
permafrost areas (Zhang and Frauenfeld et al., 2005, Lemke and Ren et al., 2007). Thus Lakselv and Karlebotn are interpreted to be on the outermost rim of palsa distribution in Northern Norway. On the contrary to the coast, the continental mires located in Finnmarksvidda undergo colder air temperatures (Figure 5 and Figure 6), and to some degree the palsa mires in Soussjavri and Aidejavri had more snow than Lakselv and Karlebotn. Thus it may be discussed that the continental palsas are, to some degree, tolerable of more snow.

Even though the palsas are favorable in low snow conditions and often have been observed without a snow cover (see Figure 21, Figure 23Figure 24 and 25), strong winds acting as redistributors of snow may also be erosive for the palsas. As Seppälä (2003) reported, the main destructive factor for in palsas in the development cycle is block erosion. And strong winter winds may work as a shaving mechanism for the steep and fractures sides of the palsas, and thus erode away the insulating peat layer (Seppälä, 2003, 2004).

Li and Pomeroy (1997) concludes that a threshold wind speed for snow transport is related to properties of the surface snow pack, such as cohesion, particle bonding and friction, which again is controlled by meteorological factors. Thus warmer temperatures lead to a denser snow pack, and larger cohesion and particle bonding, hence a higher threshold wind speed for snow transportation.

The different snow density measurements registered during winter fieldwork for this thesis may also play a part in the local environment for a palsa to form. Figure 22 shows the study sites had increasing snow density with increasing snow depth and thus a much lower snow density on the palsas than on the wet mire sites, or forested sites. This may be explained by the snow density – and snow depth relation, as snow density will increase with increasing snow depth due to the compaction by the overburden pressure (Kojima, 1967, reference therein, Lundberg and Richardson-Nälsund et al., 2006). Remembering section 2.2.1 and earlier work from Kujala and Seppälä et al., (2008) we know that the thermal conductivity increases with soil moisture. Transferring this to conductivity of water, we know that wet snow will have a larger heat exchange than that of dry snow. Allowing the cold air temperatures to penetrate further down in the snow pack and deeper into the ground, than what might have been the case for dry snow, due
to the insulating effect being greater. Figure 48 clearly represents this, as a palsa run with deep and wet snow actually yielded permafrost, while deep dry snow did not. On the contrary when the snow depth is shallow (Figure 47), the consequences on the differences in thermal conductivity of dry- and wet snow does not yield significant changes on the ALD. According to Li and Pomeroy (2010) a threshold wind speed for snow transportation is to be 8.3 m$^{-s}$ for the mean ambient air temperatures for Kautokeino meteorological station, during the period of snow cover (02.11.2014-17.05.2015) (-7.4). A total period of almost 50 days of durable wind speed is thought to be enough for the snow on top of palsas in Aidejavri to be wind transported to the lee side of the palsa, or further into the forest, as would be conclusive with the in situ snow depth measured in March (Figure 21).

However, when the snow is wet, the result might in fact be quite different. The palsa runs with wet mire snow scenarios of two different densities (200 kgm$^{-3}$ and 450 kgm$^{-3}$) showed some very interesting results (Figure 47, Figure 48), revealing that dense deep snow might actually produce permafrost while dry deep snow does not. This is in divergence of the interpretations that the study sites do not tolerate such massive snow covers, and in fact winter rain and- or warm fronts where the air temperature is above freezing may actually lead to a colder ground surface. However, as the average snow density was $\approx 300$ kgm$^{-3}$, which is substantially drier than 450 kgm$^{-3}$, one may hypothesize that a homogenous snow pack with a density of 450 kgm$^{-3}$ may not be a realistic situation for the palsa mires in Finnmark. And such high snow packs will initially lead to higher ground temperatures and thawing of the palsa mires in Finnmark.

Thin ice layers were covering several palsas in Lakselv (Figure 23 and Figure 24) and Iskorás during winter fieldwork and are possibly resulting from several freeze-thaw incidences during the winter. Such temperature alterations were in fact measured nine times during the 2014-2015 winter in Lakselv. Whereof, the warmest temperature reached above 5 °C in mid-January. Temperature peaks above freezing were also measured three times during the same winter in Karlebotn, and the result was most likely the same; creating a thin layer of ice on top of the palsas. Such freeze-thaw incidences may be enhancing the vertical flow downwards of the cold air temperatures, as ice has higher thermal conductivity than snow. On the other hand, the ice lens may as well be a larger insulator of the internal heat flowing upwards from the palsa too,
leading to warmer temperatures within the palsa. But since the ice lens was no more than 1-2 cm thick it is suggested not to be a sufficient enough insulator.

4.1.1 Conceptual model for palsa degradation

A theoretical illustration on a palsa's cyclic evolution by Seppälä and Kujala (2009) have been presented in Figure 2, and the results presented in Chapter 5 concludes with this Figure, as both an embryo palsa was observed (Figure 15 and Figure 16) and vast areas of all the palsa mires had thermokarst ponds (Figure 10, Figure 17, Figure 18, Figure 19, Figure 26, Figure 27, Figure 29, Figure 31 and Figure 32). CryoGrid 2 is able to present the ground temperatures based on laws of heat flow, however it may not be able to explain the degradation of the palsas other than the controlling factors of snow, temperature and stratigraphy. And as Figure 34 and Figure 37 shows, the difference in MAGST between a palsa and the adjacent wet mire is considerable, as much as roughly 3 °C. Signifying a large temperature gradient in the lateral direction, from a palsa to the adjacent wet mire which is hypothesized to reach the palsa and ground ice. This section will present a hypothetical and very simplified example of palsa degradation based on the energy transfer from a thermokarst lake to an adjacent palsa.

An illustration of the observed typical palsa and wet mire scenario is presented in Figure 44 and Figure 45, based on the earlier presented field work observations (see e.g. Table 1, Figure 20, 22, Figures 24 to 33, 35 and 37). The figures clearly show the general snow scenarios ranging on a palsa and in the adjacent mire and palsa lee side. As discussed, the palsas are often located right next to- or even in direct contact with water bodies which is interpreted to be the result of a degraded and thawed palsa. Thus these water bodies are interpreted to be thermokarst ponds. Large differences between palsa summit and surrounding areas both in vegetation, snow cover and temperature have been observed, as presented in Chapter 5.2, for example in the orthophoto from Iskorás (Figure 26 A-F), Soussjavri (Figure 29 A-B) and Aidejavri (Figure 32 B-D).

Because the palsa is supersaturated with of ground ice, any flow of warm temperatures towards the palsa could lead to fatal consequences for the permafrost. Due to the low density of ice, a thawing of the ground ice would result in a decrease of the palsa height and further the lateral extent. A thawing of the ground ice may possibly result in an
increased area of thermokarst lakes in the mire (Luoto Seppälä, 2003, Christensen and Johansson et al., 2004, Kirpotin and Polishchuk et al., 2009). As informed in Chapter 1.1, degradation of peatlands and the increased areas of thermokarst ponds and later drought of these ponds have been thoroughly reported for the northern hemisphere (Gorham, 1991, Laberge Payette, 1995, Payette and Delwaide et al., 2004). Interestingly, the thermokarst lakes have decreased in areal extent in discontinuous permafrost zones, while in continuous permafrost zones, the area have increased (Kirpotin and Polishchuk et al., 2009). Indicating the indicating both thawing of the segregated ground ice and drainage of the lakes in discontinuous permafrost areas, and areal decrease of the thermokarst lakes related to the increase in air temperature and decrease in excess ground ice. Therefor the geocryological transformation described by Kirpotin and Polishchuk et al. (2009) is highly important, when such trends lead to an accelerated climate warming, as have been reported by numerous of authors (Cao and Gregson et al., 1998, Christensen and Johansson et al., 2004, Walter and Zimov et al., 2006, Jackowicz-Korczyński and Christensen et al., 2010).

With the vast lateral degradation reported for northern Norway (Borge and Westermann et al., 2016), Sweden (e.g. Zuidhoff and Kolstrup, 2000, Åkerman and Johansson, 2008), Canada (e.g. Laberge and Payette, 1995, Tremblay and Bihry et al., 2014,), one may assume this to be a warning on the future of the much more extensive wetland permafrost in the Russia which have also been reported on degradation (Pavlov, 1994, Walter and Zimov et al., 2006, Kirpotin and Polishchuk et al., 2009, Kirpotin and Polishchuk et al., 2011) and emission of greenhouse gasses from the thawing organic layer and thermokarst lakes hemisphere. As informed above, and presented in Figure 2, when a palsa is in the degrading phase and in contact with water, the degradation will escalate.

In order to better understand what happens when the ground ice in a palsa thaws and the palsas are in direct contact with thermokarst lakes, or in the adjacent vicinity of it, a conceptual model may be presented.
Figure 44: Illustration over the conceptual model for palsa degradation by the thermal influence of the adjacent water body. Red arrow indicates direction of heat flow from the warmer water body to the adjacent palsa and ground ice. Notice the distinct green vegetation on surrounding areas of the mire, while other vegetation reigns on the palsa itself. The wet mire – and palsa scenario is also included in the winter scenario with height scale to the left. Distance from thermokarst lake to palsa is only 20 cm in this scenario.
The thermokarst ponds are measured to have MAGST temperatures above freezing point, due to the significant insulating properties of snow. From this one would insinuate that the ponds are unfrozen during the winter, as is the situation in Iskorás (Figure 37), and hence warmer than the adjacent palsa which has very little snow. From thermodynamics we know that the direction of heat flow will always be from warm medium towards colder adjacent medium. Transferring this to the palsa and peat plateau environment, one would assume a heat flow from wet mire sites to palsas. From Fourier’s law of heat conduction integrated for a homogenous, material of one dimensional, 1D, geometry, Eq. 11 is presented, which is applied for calculating the potential total energy flowing from the thermokarst lake to the adjacent palsa.

The MAGST for the wet mire and palsa is presented in Figure 45 and 46, and represents the GST-loggers with ground surface temperature from August 2014-August 2015. MAGST palsa and wet mire equals 274.3 K and 277.2 K, respectively. The temperature difference (ΔT) thus equals ≈ 3 K. And from the 0.2 m distance from palsa to wet mire (Δd) the lateral heat flow is calculated, by knowing the thermal conductivity of water to be an estimated value of 0.6 Wm⁻¹K⁻¹, for a mean water temperature of 274.3 K, following Huber and Perkins et al. (2012):

\[ F = -k \frac{\Delta T}{\Delta d} = -0.6 \frac{W}{mK} \times \frac{3 K}{0.2 m} = -9 \frac{W}{m^2}. \]  

(11)

Knowing 1 W = 1 J/s, we can estimate how much energy does the palsa receive from the adjacent thermokarst lake per year, see equation 12, and thereof the potential annual loss in height of the palsa ground ice (equation 13) by knowing the latent heat of fusion of water (334 MJm⁻³ (Dickinson and Harper et al., 1903)), as introduced in equation 2 in Chapter 4.2.2.

\[ -9 \frac{W}{m^2} = -9 \frac{J}{m^2 s} \rightarrow -9 \frac{J}{m^2 s} \times 3600 \times 24 \times 365 = -2.84 \times 10^8 \text{ Jm}^{-2}. \]  

(12)

Thaw depth of ice pr. year = \( \frac{-284 \, 000 \, 000 \, \text{Jm}^{-2}}{334 \, 000 \, 000 \, \text{Jm}^{-3}} = -0.85 \, \text{m} \)

(13)

In fact, with the 2014-2015 climate forcing, this simplified calculation insinuates that 0.85 m of the ground ice will thaw with the specified distance of palsa and wet mire in
one year. This would result in a massive decreasing of the ground ice within a short time period. However, the peat layer may be covering the ground ice from the estimated lateral heat flux, and thus the distance increases. Also, the flow of warm air from the thermokarst lake will not only in the lateral direction flow towards the palsa, but also to the other adjacent mire areas, and in vertical direction towards the air during summer, or towards the snow pack during winter.

However, what would the total energy flux be when the distance from the ground ice increases? Figure 45 illustrates a degrading palsa during summer and winter, with the same GST-logged MAGST’s as in the example above (Figure 44). This palsa is also in the adjacent area of a thermokarst pond, however this scenario exemplifies a larger distance from the palsa to the pond.

The lateral heat flux decreases with increasing depth from the water pond. Therefore, with a distance of 5 meters between the palsa and water body, as illustrated in Figure 45, a total height of 3.4 centimeters may be thawed by the heat flux from the water pond with the logged MAGST. This is summed up in equation 16 by following equation 14 and 15.

\[
F = -k \frac{\Delta T}{\Delta d} = -0.6 \frac{W}{mK} \times \frac{3 K}{5 m} = -0.36 \frac{W}{m^2} \\
\]  
(14)

\[
0.36 \frac{J}{m^2s} \times 3600 \times 24 \times 365 \approx -1.14 \times 10^7 Jm^{-2} \\
\]  
(15)

Thaw depth of ice pr. year = \[\frac{-114000000 Jm^{-2}}{334000000 Jm^3} \] = -3.4 cm  
(16)
Figure 45: Illustration over the conceptual model for palsa degradation by the thermal influence of the adjacent water body, located 5 meters from the palsa. Notice the direction of heat flow from the warmer thermokarst lake. See Figure 44 for legend.
In the two presented scenarios (Figure 44 and Figure 45) one is making simplified assumptions, such as assuming the water pond to be in 1-D, and thus not in movement as one may support by the presence of vegetation on the ponds which keeps the water still. However, when applying this scenario to a 2D transient thermal model, as e.g. Myhra and Westermann et al. (2015) have done in investigating permafrost in rock walls in relation to climate sensitivity, a more realistic scenario is obtained. This is because two dimensional models include parameters dependent on both time and space, and the latent heat of ice. Thus a 2D model might have been able to clarify the lateral heat exchange more thoroughly. Nonetheless, the simplified 1D scenarios yield assumptions on what might be the case for energy fluxes when the snow depth varies as much as 60 cm between a palsa and an adjacent wet mire point.

But if one imagines the water pond is no longer homogenous and in 1D simulations, the ground ice thaw-rate situation would be further escalated, as the thermal conductivity of the thermokarst lake would have inter annual variations with snow cover. Remembering section 2.4.3, the thermal conductivity of ice is higher than of ice, actual four times higher following Goodrich (1978) and Smith and Risebourgh (2002). Then the conductivity ratio would increase for the thermokarst pond, and the latent heat flux would also increase, which would result in an even deeper thawing of the ground ice in the palsa.

### 6.3 Model results

Transient thermal modeling of palsas has not been performed earlier for northern Norway, and neither have transient thermal modeling of palsas with in situ soil and climate forcing. The results are surprisingly well fitting with the present ALD for the selected palsas in Finnmark (Figure 34Figure 35Figure 36Figure 37Figure 38,). The presented palsa stratigraphy and palsa snow forcing seem to reproduce the current defining limits for palsa mires in Finnmark (Figure 43) as the majority of the mapped palsas are delimited by the -5°C to -3 °C isotherm, which Borge and Westermann et al. (2013) also projected.

The grid based transient thermal model CryoGrid 2 has been forced with different stratigraphy- and snow scenarios on both a small- and regional scale for palsa mires in
Finnmark. The results show that the ALD for a typical palsa in Finnmark is in good agreement with the situ ALD measurements (≈ 45 – 50 cm) when assuming the snow depth is in consistence with field measurements (≈ 10 cm). This shows that the stratigraphy and snow scenario presented in Table 1 and Figure 9, respectively, have been proven to be a realistic scenario for palsa mires in Finnmark. From Figure 42 it seems that the palsas in Finnmark, especially in coastal areas, are dependent on a very low snow cover in order to sustain the present air temperatures, as have been discussed above. Thus the palsa mires are reliant on sufficient wind (≈5 ms⁻¹), in order for the winter precipitation to be transported away from the palsa summits. Wind have been reported to be a controlling factor for palsa formation (Seppälä, 1982), and this seems to be the case for all palsas in Finnmark, as only low snow cover produces the present palsa distribution for the study area. However as the snow tend to accumulate on lee sides of palsas and in the general adjacent areas of the permafrost core, the temperature regime may still undergo degradation in the present climate. This is clearly demonstrated in Figure 41, as wet mire scenario does not produce permafrost temperatures. Thus, a deep snow cover during the winter may have fatal consequences for a palsa, as have been reported in earlier studies (Zuidhoff and Kolstrup, 2000; 2005, Zuidhoff, 2000, Jean and Payette, 2014, Christensen and Johansson et al., 2006).

The two isoline delimiting the ground temperatures from -5 to -4°C and -4 to -3 °C covered almost all of Finnmark, except the coastal areas. From this one may insinuate that if the topography in Finnmark were more or less consisting of mires, the palsa distribution would have covered almost 150 km². As this is not the case, the aerial extent of permafrost in mires is noticeably less, and thus CryoGrid 2 substantially overestimates the permafrost area of mires in Finnmark. Earlier grid based modeling for Finnmark have insinuated that permafrost in mires could cover an area of several thousand square kilometers (Farbrot and Isaksen et al., 2013, Gisnås and Etzelmüller et al., 2013). Gisnås and Etzelmüller et al. (2013) modeled permafrost for the mountainous areas of Norway in relation to small scale variability of snow, which is similar to what have been done in this thesis. However the mapped palsa distribution by Borge and Westermann et al. (2016), which is presented in Figure 43, shows that the two former authors and overestimated the current situation, as the areal extent of permafrost in mires were mapped 110 km².
The active layer depths produced by CryoGrid2 seems to be in consistence with the present situation in all of the sites, however the climate forcing used for the model might not be representative for the years to come with the observed climate change and predicted future climate warming and especially since the largest temperature increasing have been observed in higher latitudes (IPPC, 2014, Christiansen et al., 2010). CryoGrid 2 have been proven to be a powerful tool to infer temperature regime of the ground according to conductive heat transfer in the soil and in the snow pack in a changing climate. When inferring the model with field obtained measurements and parameters the model reproduces the active layer measurements for all of the study sites (Figure X) and is consistent with the mapped palsa mire distribution for Finnmark, and shows that palsas are restricted mainly to the inner parts of Finnmark, at Finnmarksvidda where the air temperatures are colder than at the coast.

In contrast to the equilibrium permafrost model of Gisnås and Westermann et al. (2015), which facilities statistical representations of sub-grid variability of snow depth, CryoGrid 2 does not assume small scale variability of snow; it assumes a steady state homogenous snow pack of the height and duration as determined in the initial forcing, see Figure 5, Figure 8, Figure 9 and Appendix Nonetheless, it reproduces realistic ALD and ground temperatures for the present climate in a steady state with the assumptions of a homogenous snowpack representative for the current snow scenarios in palsa mires. As discussed on section 6.1.2. the height and density and of the snow pack are crucial factors for mire permafrost conditions, thus this assumption is a factor of error for the modeled results.

Similar transient thermal modeling for Northern Norway have been performed by other authors, e.g. Farbot and Isaksen et al. (2013), and have also been proven to be powerful tools in modeling the ground thermal regime. The latter author modeled ground temperatures using a transient heat flow model with senorge climate forcing and validated the results against time series of borehole measurements. The surface offset values for the five study sites, presented in Table 2, shows the difference between MAGST and MAGT, in this thesis the MAGT is estimated to be at the thaw depth. Which might be too low and thus yielding smaller surface offsets than what might be the current situation The surface offset values were ranging from -0.27 °C to 0.26 °C for the palsa scenario (Table 2). In comparison, surface offset from 4 boreholes in Iskoras for
2007-2009 were presented by Farbrot and Isaksen et al. (2013) and were ranging from roughly 0 to 3 °. These bore holes are located on Northern and Southern side of the Iskorás mountain on low vegetated ground. However, the borehole with ~ 0°C surface offset was located on the Northern side of Iskoras mountain, by the foot of the mountain and thus not very far from the palsa mire (~2 km). From this one may insinuate that the ground temperatures produced in CryoGrid 2 is in consistence with in situ measurements in Iskorás.

Farbrot and Isaksen et al. (2014) describes Finnmarksvidda to be in a zone of permafrost warming. The field observations are supporting this by the extensive lateral erosion (e.g. Figure 10-4, Figure 19, thermokarst lakes (Figure 17 and Figure 18), deep ALD on some palsa (e.g. Figure 24-B, F) and snow accumulation in depressions on the palsa (e.g. Figure 24-C,E). However, the model results (Figure 25) insinuates that permafrost in mires may sustain if the temperature forcing for 2014-2015, and the snow depth on the palsa was representative for the future climate scenario. As both temperature and snow depth is estimated to increase in the near future(Osterkamp Romanovsky, 1999, Burgess and Smith et al., 2000, Brown Romanovsky, 2008, Christiansen and Etzelmüller et al., 2010, Romanovsky and Smith et al., 2010).
6.4 Suggestions for further research

Based on the results and interpretations from this thesis, I encourage further investigation on the following topics:

- The state of the arts permafrost models are hampered by ambiguities in soil stratigraphy, vegetation and land cover. However, the presented soil stratigraphies may provide a broader accuracy on modeling permafrost in peatlands. From this I encourage further investigation on modeling permafrost in regions of discontinuous permafrost with respect to the presented palsa scenario.

- Perform systematic snow and ALD measurements in successive years to come on the same study site points, in order to conclude on the evolution of the palsa mires in Finnmark, and investigate the rate of degradation with the trend in climate warming.

- Analyze the interior of the palsa and the depth of the excessive ground ice with GPR over successive years, which will yield a sufficient deep soil stratigraphy as well as the depth and development of the ground ice.

- Incorporate GST-logger data from a representative selection of palsa mires in northern Norway into CryoGrid 2, which would yield a sufficient dataset to give a further validation of the stratigraphies and snow scenarios as presented in this thesis.
7 Conclusions

Permafrost models are efficient and strong tools in inferring the ground thermal regimes, and research on this topic are numerous and increasing. However, such models have inaccuracies in the local topological aspects incorporated in the simulations, such as soil, snow distribution, vegetation and organic content, which are crucial factors for permafrost distribution. This thesis presents the importance of estimating these local factors governing on palsa mires when modeling permafrost distribution.

The transient thermal model, CryoGrid 2, have proven to be a powerful tool in permafrost modeling of palsa mires in northern Norway, which demarcate the westernmost extent of the much larger peat plateaus in western Siberia. With the recently reported substantial permafrost and peatland degradation in the northern hemisphere follows hazardous consequences for the climate – permafrost feedback which are hard to predict and have not been fully understood. Thus further investigation on this field is crucial for understanding the future magnitude and outcome of permafrost degradation and climate warming.

The modeling of palsa mires in northern Norway have proven that the palsa distribution within coastal areas of Finnmark seem to be controlled by a seasonally low snow cover. Moreover, these landforms which form the outermost limits of discontinuous permafrost in Scandinavia will not survive a further degradation of the insulating peat layer, and as a consequence greenhouse emissions will most likely accelerate.

Even though the results from the model run and field work is in consistence with one another, and with former published work on the subject, further research is motivated in stratigraphic investigations, in situ measurements of snow and active layer depth which may improve further modeling of permafrost in organic rich soil. And the modeled results should also be correlated with boreholes in Finnmark.

Moreover, the final results and conclusions are listed below:

- Based on field work during summer and winter, and analysis of soil samples from two palsa mires, three representative soil stratigraphies have been presented, palsa, wet mire and embryo palsa. The palsa stratigraphy is comparable with other
palsa mire stratigraphies presented for Sweden and Canada. These mire stratigraphies have been utilized as soil forcing in CryoGrid 2.

- CryoGrid 2 is able to reproduce the present active layer depths and ground surface temperatures for palsa mires in Finnmark, northern Norway using soil and snow forcing obtained from field work, and air temperature forcing from met stations. Thus, the presented soil forcing is interpreted to be describing one of the local factors for permafrost distribution in northern Norway.

- Snow depth, soil stratigraphy and wind seem to be the controlling factors for palsa mires in Finnmark. The coastal palsa mires are dependent on very low snow depth (< 10 cm) in order to survive with the present climate, and continental sites are more tolerable of snow. However the reported massive palsa degradation will most likely continue and accelerate in the near future, due to both climate warming and the abundant thermokarst lakes in the mires which accelerate thawing of the palsas ground ice.

- A conceptual model for palsa degradation have been presented, which is motivated by the lateral heat transfer by thermokarst ponds towards palsas, due to the measured MAGST difference between palsa and thermokarst pond. This simple estimate is underlining the thawing effect of thermokarst ponds in the adjacent areas of the palsa. The thawing effect is hypothetically increasing with decreasing distance between the palsa and the thermokarst lake.

- Snow and active layer depth seem to be in correlation with one another in palsa mires, as snow depth is deepest on sites without permafrost, and less than 20 cm for sites with permafrost.

- Systematic measurements on soil wetness and active layer depth signalize that no permafrost is present in wet mire soil. When initializing CryoGrid 2 with wet mire scenario, no permafrost is produced. Soil wetness increase with decreased elevation within the palsa mires.

- Thaw depth on palsas in Finnmark is measured to be roughly averaged to 50 cm, and is more or less the same for coastal and continental mires even though the climate is extremely divergent from the coast to the continental parts of the county. This is interpreted to be explained by the 50 cm deep peat layer which is insulating the ground ice from the summer warmth, and permits the outward
flow of the warmer ground temperatures during winter and due to the wind drift of snow from the palsa summits during winter.

- Thaw depth for palsa mires in Finnmark is presented and compared to modeled thaw depths with the presented soil and snow scenarios. The results are in good agreement with in situ measurements and prove that CryoGrid 2 may reproduce the current state of the palsa mires with in situ climate and soil forcing.

- A sensitivity analysis on the soil stratigraphy proved that no permafrost is produced with deep snow depth; this was interpreted to be due to the extensive insulating effect of the snow on the ground surface temperatures.

- The modeled palsa ground temperatures is also in consistence with earlier reported palsa distribution map in Finnmark by Borge and Westermann et al., (2016), and concludes that palsas in Finnmark are located in areas of ALD of 40-50 cm with 1-m average annual temperatures within the ranges of -5 °C to -3 °C.
Literature


Blyakharchuk, T. A. & Sulerzhitsky, L. D. (1999). Holocene vegetational and climatic changes in the forest zone of Western Siberia according to pollen records from the extrazonal palsa bog Bugristoye. The Holocene, 9, 621-628.


Zhang, Y. C., Wenjun


Åkerman, J. (1980). *Studies on periglacial geomorphology in West Spitsbergen*, Royal University of Lund, Department of Geography.
Appendix

Additional figures and Tables

In situ active layer depth – Lakselv

*Figure 46: Stacked histogram presents soil condition (wet/dry) with measured ALD - and snow depth during 13th of March 2016 at Lakselv study site (Oksebergmyra).*
Table 2: Overview of results from palsa runs with Senorge snow and palsa snow. Modeled and measured ALD are more or less comparable with one another, only Lakselv seems to deviate some. The ALD presented in Figure 39 was modeled with palsa- and senorge snow with a snow density of 300 kgm⁻³, while in the regional model run, snow density was adjusted after the winter field work observations to a density of 200 kgm⁻³, as was in consistence with the snow density on top of the palsas (see Figure 22) and literature (Lundberg and Richardson-Näslund et al., 2006). N-factors used for the initialization were ranging from 0.6 for palsa runs and 0.1 - 0.2 for senorge and wet mire snow runs for the study sites. N-factor of 0.6 was in consistence with in situ GST and air temperature logged for Iskorás. However, for the regional model run the n-factor is implemented in the run, and one does not need to estimate it manually.

<table>
<thead>
<tr>
<th>Study Site</th>
<th>In situ AL Depth</th>
<th>Modeled AL Depth</th>
<th>TTOP</th>
<th>MAAT</th>
<th>MAGT</th>
<th>MAGST</th>
<th>Thermal Offset</th>
<th>Surface Offset</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Senorge Snow Scenario (ρ=300kgm⁻³)</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Iskorás</td>
<td>0.40-0.70</td>
<td>0.55</td>
<td>-0.25</td>
<td>-0.35</td>
<td>-0.48</td>
<td>0.1</td>
<td>-1.44</td>
<td>-1.31</td>
</tr>
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<td>Karlebotn</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.88</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Lakselv</td>
<td>0.55-0.80</td>
<td>-</td>
<td>-</td>
<td>0.69</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Aidejavri</td>
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<td>0.55</td>
<td>-0.38</td>
<td>-0.93</td>
<td>0.37</td>
<td>-1.99</td>
<td>-2.35</td>
<td>-2.01</td>
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<tr>
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<td>0.48</td>
<td>0.49</td>
<td>-1.26</td>
<td>-0.62</td>
<td>-1.95</td>
<td>-0.89</td>
<td>-1.09</td>
<td>0.25</td>
</tr>
<tr>
<td><strong>Palsa Snow Scenario (ρ=300kgm⁻³)</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Iskorás</td>
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<td>-1.34</td>
<td>-0.35</td>
<td>-1.81</td>
<td>-0.08</td>
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<td>-0.70</td>
<td>1.13</td>
<td>-1.83</td>
<td>-0.25</td>
</tr>
<tr>
<td>Lakselv</td>
<td>0.55-0.80</td>
<td>0.51</td>
<td>-0.53</td>
<td>0.69</td>
<td>-0.50</td>
<td>1.07</td>
<td>-1.51</td>
<td>0.14</td>
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<tr>
<td>Aidejavri</td>
<td>0.40-0.75</td>
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<td>-1.90</td>
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<td>0.15</td>
<td>-2.41</td>
<td>-0.26</td>
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<tr>
<td>Soussjavri</td>
<td>0.48</td>
<td>0.51</td>
<td>-0.62</td>
<td>-2.31</td>
<td>-0.36</td>
<td>-1.95</td>
<td>-0.19</td>
<td></td>
</tr>
<tr>
<td><strong>Wet mire snow Scenario (ρ=300kgm⁻³)</strong></td>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Iskorás</td>
<td>0.40-0.70</td>
<td>NP</td>
<td>-</td>
<td>-0.35</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Karlebotn</td>
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<td>NP</td>
<td>-</td>
<td>0.88</td>
<td>-</td>
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<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Lakselv</td>
<td>0.55-0.80</td>
<td>NP</td>
<td>-</td>
<td>0.69</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Aidejavri</td>
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<td>-0.58</td>
<td>1.12</td>
<td>-1.69</td>
<td>-1.23</td>
</tr>
<tr>
<td>Soussjavri</td>
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<td>-0.23</td>
<td>-0.62</td>
<td>-1.44</td>
<td>-0.68</td>
<td>-0.76</td>
<td>-0.25</td>
</tr>
</tbody>
</table>
Figure 47: Presenting the modeled active layer depth for the study sites for palsa run with different density of palsa snow scenarios. Notice that the ALD does not change significantly with dry snow ($p = 200 \text{ kgm}^{-3}$) and dense snow ($p = 450 \text{ kgm}^{-3}$) when the snow depth is 10 cm. See legend.

Figure 48: Presenting the modeled active layer depth for the study sites for a palsa run with different densities of wet mire snow scenarios. Notice that very dense wet mire snow ($p = 450 \text{ kgm}^{-3}$) actually yields permafrost, while dry snow ($p = 200 \text{ kgm}^{-3}$) does not. See legend. PF = permafrost.
Meteorological Data

A

Wind Velocity (ms⁻¹)

- >20.2
- 15.3-20.2
- 10.3-15.2
- 5.3-10.2
- 0.3-5.2

Stille (%)

B

Wind Velocity (ms⁻¹)

- >10
- 7.6-10
- 5.2-7.6
- 2.8-5.1
- 0.3-2.7

Stille (%)

Figure 49: Wind-rose diagrams presenting the measured wind velocities and predominant wind direction during 1961-1990 for (A) Karlebotn (Kirkenes Met Station) and (B) Lakselv (Banak Met Station).
Figure 50: Presenting the mean annual normal in temperature and snow depth for the normal period 1971-2000. Map downloaded at senorge.no/?p=climate, and modified by the author. Data available by The Norwegian Water Resources and Energy Directorate (Engeset and Tveito et al., 2004a, 2004b, Mohr, 2008, Mohr Tveito, 2008).
Matlab® Script of CryoGrid 2

%Before cryoGrid2.m can be run, the C-file "testMex.c" must be compiled by typing "mex testMex.c" in the command prompt
%This will only work if a C compiler is installed, check by using "mex -setup"
%If no Compiler is found, a C Compiler must be downloaded.
%C compiler used for CryoGrid 2 on a MacBooc Pro OSX Yosemite 10.10.5  Macontosh HD 2.5 GB 1600 mHZ was X Code version 5.0.2.

Q=0.05; % heat flux at lower boundary [ W/m2]
rho_snow=200; % density of the snow [kg/m3] 200 = dry, 300 = med. 450 = wet
kh_snow = 2.2.*(rho_snow./1000).^1.88; % Yen (1981)
k_bedrock=3; % thermal conductivity of the mineral component of the soil [W/mK]=3

% THE SOIL STRATIGRAPHY FORCINGS:
% soil stratigraphy - column 1: depth [m];
% column 2: volumetric water content
% column 3: volumetric mineral content
% column 4: volumetric organic content
% column 5: not used
% column 6: code for soil type: 1: sand, 2: silt
% column 7: not used

%PALSA Stratigraphy
soilParam=[-5 0.3 0.05 0.15 0 2 0; 0.44 0.3 0.05 0.15 0 2 0;... %Organic Layer
0.46 0.8 0.05 0.15 0 1 0; 1.95 0.8 0.05 0.15 0 1 0;... %Ground Ice Layer
2.05 0.5 0.5 0 0 2 0; 14.9 0.5 0.5 0 0 2 0;... Mineral Layer
15.1 0.03 0.97 0.0 0.05 1 0; 5000 0.03 0.97 0.0 0.05 1 0]; %Bedrock

%Thermokarst Lake:
soilParam=[-5 0.8 0.05 0.15 0 1 0; 1.10 0.8 0.05 0.15 0 1 0;... %Water Body with mire vegetation
1.12 0.5 0.5 0 0 2 0; 14.9 0.5 0.5 0 0 2 0;... %Mineral Layer
15.1 0.03 0.97 0.0 0.0 1 0; 5000 0.03 0.97 0.0 0.0 1 0]; %Bedrock

% Embryo Palsa
soilParam=[-5 0.25 0.05 0.15 0 1 0; 0.2 0.25 0.05 0.15 0 1 0;... %Organic Layer
0.21 0.8 0.05 0.15 0 1 0; 1.94 0.8 0.05 0.15 0 1 0;... % Ground Ice Layer
2.05 0.5 0.5 0 0 2 0; 14.9 0.5 0.5 0 0 2 0;...% Mineral Layer
15.1 0.03 0.97 0.0 0.0 1 0; 5000 0.03 0.97 0.0 0.0 1 0]; %Bedrock
The model is adjusted for the snowfall dates and the build-up of the snow cover. The second column is the height in meters. The first date MUST be before the first date specified by your 1 year forcing time. Snow Series obtained from in situ snow measurements of from climate station. Listed for each study site. Date of time series and the last, and snow height.

Kautokeino Snow Forcing

```matlab
date_struct_snow = [{'17.08.2014',0; '20.08.2014', 0; '22.10.2014', 0.02; '25.10.2014', 0.03; '26.10.2014', 0.00; '04.11.2014', 0.12; '11.11.2014', 0.25; '21.11.2014', 0.23; '03.12.2014', 0.29; '04.12.2014', 0.30; '11.12.2014', 0.26; '16.12.2014', 0.32; '21.12.2014', 0.36; '31.12.2014', 0.40; '05.01.2015', 0.44; '11.01.2015', 0.50; '18.01.2015', 0.58; '04.02.2015', 0.62; '07.02.2015', 0.58; '09.02.2015', 0.54; '26.02.2015', 0.52; '01.03.2015', 0.49; '11.03.2015', 0.47; '31.03.2015', 0.49; '06.04.2015', 0.44; '08.04.2015', 0.40; '15.04.2015', 0.45; '22.04.2015', 0.34; '25.04.2015', 0.30; '09.05.2015', 0.20; '11.05.2015', 0.18; '12.05.2015', 0.15; '15.05.2015', 0.12; '16.05.2015', 0.10; '18.05.2015', 0.05; '21.05.2015', 0; '22.05.2015', 0; '01.09.2015', 0};
```

Karlebotn Snow Forcing

```matlab
date_struct_snow = [{'18.08.2014',0; '20.08.2014', 0; '1.11.2014', 0.0; '31.12.2014', 0.25; '23.02.2015', 0.47; '19.03.2015', 0.34; '20.03.2015', 0.60; '24.03.2015', 0.49; '11.04.2015', 0.37; '13.03.2015', 0.60; '19.04.2015', 0.23; '23.04.2015', 0.13; '04.05.2015', 0.02; '05.05.2015', 0.0; '31.08.2015', 0; '01.09.2015', 0};
```

ISKORAS Snow Forcing

```matlab
date_struct_snow = [{'01.01.2014', 0; '28.10.2014', 0.05; '13.11.2014', 0.20; '04.12.2014', 0.25; '30.12.2014', 0.30; '10.01.2015', 0.40; '22.02.2015', 0.50; '28.02.2015', 0.55; '13.03.2015', 0.50; '19.04.2015', 0.47; '23.04.2015', 0.22; '10.05.2015', 0.22; '21.05.2015', 0.22; '22.05.2015', 0; '01.09.2015', 0};
```
%Soussjavri Snow Forcing
date_struct_snow=
{ '18.08.2014', 0;
'18.10.2014', 0.02;
'27.10.2014', 0.05;
'28.10.2014', 0.02;
'29.10.2014', 0.07;
'06.11.2014', 0.13;
'15.11.2014', 0.15;
'04.12.2014', 0.27;
'31.12.2014', 0.32;
'08.01.2015', 0.40;
'31.01.2015', 0.42;
'06.02.2015', 0.55;
'10.02.2015', 0.57;
'28.02.2015', 0.48;
'07.03.2015', 0.44;
'19.03.2015', 0.42;
'24.03.2015', 0.55;
'04.04.2015', 0.55;
'09.04.2015', 0.50;
'15.04.2015', 0.52;
'22.04.2015', 0.36;
'01.05.2015', 0.36;
'03.05.2015', 0.40;
'08.05.2015', 0.34;
'09.05.2015', 0.22;
'12.05.2015', 0.18;
'15.05.2015', 0.05;
'21.05.2015', 0.01;
'23.05.2015', 0;
'01.09.2015',0 };
% Wet Mire Snow Scenario
date_struct_snow=
{ '17.08.2014',0;
'20.08.2014',0;
'14.10.2014', 0.05;
'18.10.2014', 0.10;
'30.10.2014', 0.10;
'17.11.2014', 0.10;
'23.11.2014', 0.10;
'01.12.2014', 0.10;
'21.12.2014', 0.60;
'24.12.2014', 0.60;
'27.12.2014', 0.60;
'02.01.2015', 0.60;
'08.02.2015', 0.60;
'17.02.2015', 0.60;
'01.03.2015', 0.60;
'11.04.2015', 0.60;
'12.04.2015', 0.60;
'21.04.2015', 0.60;
'23.04.2015', 0.60;
'04.05.2015', 0.60;
'09.05.2015', 0.10;
'15.05.2015', 0.04;
'20.05.2015', 0.04;
'21.05.2015', 0.00;
'31.08.2015', 0;
'01.09.2015',0 };%

% Palsa Snow Scenario 10 cm snow
date_struct_snow=
{ '17.08.2014',0;
'20.08.2014', 0;
'01.11.2014', 0.0;
'31.12.2014', 0.10;
'23.02.2015', 0.10;
'19.03.2015', 0.10;
'20.03.2015', 0.10;
'24.03.2015', 0.10;
'11.04.2015', 0.10;
'13.03.2015', 0.10;
'19.04.2015', 0.10;
'23.04.2015', 0.10;
'04.05.2015', 0.02;
'05.05.2015', 0.0;
'01.09.2015',0
};%

%Downscaled Snow Scenario
date_struct_snow=
{ '01.01.2014', 0.0;
'12.10.2014', 0.0;
'13.10.2014', 0.01;
'16.10.2014', 0.03;
'17.10.2014', 0;
'28.10.2014', 0.05;
'13.11.2014', 0.20;
'04.12.2014', 0.25;
'30.12.2014', 0.30;
'10.01.2015', 0.40;
'22.02.2015', 0.60;
'28.02.2015', 0.60;
'13.03.2015', 0.60;
'19.04.2015', 0.60;
'23.04.2015', 0.60;
'07.05.2015', 0.20;
'21.05.2015', 0.0;
'01.09.2015',0 };
grid=[[0:0.02:1] [1.1:0.1:3] [3.2:0.2:5] [5.5:0.5:10] [10.5:0.5:20] [21:1:30] [35:5:50] [60:10:100]]; arraySizeT=1002; % number of points in the lookup tables for conductivity and capacity - should be at least 1000

% three column vectors of equal length are required for the forcing data - time (in Matlab time, i.e. % daily increment of 1), T_forcing (in degree C), and snowHeight_forcing (in m)

% The study site variable is defined below.
% numberOfYears is how often it gets repeated for spin-up % range is the range of the data one wants to process, corresponds % to one year, could also be e.g. 1:2192 if one has 6 measurements per day for 1 year. % rk is from the TTOP model (Gisnås et al. 2013), which is now used to initialize the very first steady-state.

targetVariable=Studysite; % Soussjavri; Iskoras, Aidejavri; % Lakselv; % Karlebotn % GST Palsa % GST Wet Mire
targetVariableSnow=generateSnow(date_structure_snow, targetVariable(:,1));
numberOfYears=5; % Number of years for the equilibrium conditions are reached
range=1:366; % Size of the air temperature forcing data, e.g. 365 equals 1 measurements per day.
rk=0.6;
nf=0.4; % we now need to introduce an n-factor for snow
% The n-factor has to be adjusted for every run, so that the temperature e.g. at 2m % depth is not changing too much over the initialization-period. In general, little % snow will be nf=0.9, medium snow around 0.5-0.6, and a lot of snow (>1m) about 0.2;

% Initiation the arraysize of time span and snow depth forcing for the % selected number of years of model initialization
adjustmentTime=squeeze(repmat(-numberOfYears+1:0, range(end), 1));
adjustmentTime=adjustmentTime(:)*365;
time=repmat(targetVariable(range,1), numberOfYears, 1)+adjustmentTime;
T_forcing=repmat(targetVariable(range,2), numberOfYears, 1);
snowHeight_forcing=repmat(targetVariableSnow(range,2), numberOfYears, 1);

% do not modify from here onwards

day_sec=24*3600;
time=[time(1,1):time(2,1)-time(1,1):time(1,1)+(size(T_forcing,1)-1)*time(2,1)-time(1,1):time(1,1))];
t_SPAN = time.*day_sec;
Tsurf=[time T_forcing];
snowHeight=[time snowHeight_forcing];
Tsurf(find(snowHeight(:,2)>0 & Tsurf(:,2)>0),2)=0; % set surface forcing to zero degrees when it is positive and there is snow
maxSnow=max(snowHeight(:,2));
gridCellSizeSnow=0.02;
%grid on which conductivity information lives (boundaries of grid cells)
K_grid = [-1*round(maxSnow/gridCellSizeSnow)*gridCellSizeSnow-0.1:gridCellSizeSnow:-gridCellSizeSnow] grid;
%grid on which capacity and temperature information lives (midpoints of grid cells)
cT_grid=(K_grid(1:end-1)+K_grid(2:end))/2;
%initialize the lookup tables for conductivity and capacity
[cT_frozen, cT_thawed, K_frozen, K_thawed, conductivity, capacity] = initialize(soilParam, arraySizeT, kh_snow, rho_snow, cT_grid, K_grid, k_bedrock);

warning off all
%now initializes with TTOP as temperature at the surface
TTOP=(nf.*nansum((Tsurf(:,2)<0).*Tsurf(:,2))+rk.*nansum((Tsurf(:,2)>=0).*Tsurf(:,2)))./size(Tsurf,1)
T0=steadyState(TTOP, Q, conductivity, cT_grid, K_grid, K_frozen, K_thawed, arraySizeT);
%initializes the temperature profile as a steady-state profile with average
%T_forcing somehow corrected for snow effect - this is only coarse initial
%guess, spin-up is required - can also be replaced by measured temperature
%profile

data.Tsurf = Tsurf;
data.snowHeight = snowHeight;
data.Q = Q;
data.cT_frozen = cT_frozen;
data.K_frozen = K_frozen;
data.conductivity = conductivity;
data.capacity = capacity;
data.arraySizeT = arraySizeT;
data.K_grid = K_grid;
data.cT_grid = cT_grid;
data.cT_lowerSnowCell = find(cT_grid(:,1)<0);
cT_lowerSnowCell=cT_lowerSnowCell(end);  %cell where snow ends
data.cT_lowerSnowCell = cT_lowerSnowCell;
data.cT_gridSizeSnow = K_grid(2,1)-K_grid(1,1);
data.sizeTsurf = size(Tsurf,1);
data.sizeSnowHeight = size(snowHeight,1);
data.sizeT = size(cT_grid,1);
RelTol=1e-3;  % relative integration tolerance
AbsTol=1e-5;  % absolute tolerance around 0 degree
options = odeset('RelTol',RelTol,'AbsTol',AbsTol);

[t,T] = ode45(@(testMex3,t_SPAN, T0,options, data) %solve the ODE
for i=1:size(t_SPAN,2)  %set grid cells above snow to NaN
    snow = interp1(snowHeight(:,1),snowHeight(:,2),t_SPAN(1,i)/day_sec);
cT_lastSnowCell = find(cT_grid(:,1)<-snow);
cT_lastSnowCell=cT_lastSnowCell(end);
T(i,1:cT_lastSnowCell)=repmat(NaN, 1, cT_lastSnowCell);
end

T=T'; %Temperature for each depth with time(cellsize defined above, in grid).