Geophysical mapping of palsa peatland permafrost

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Abstract

Permafrost peatlands are hydrological and biogeochemical hotspots in the discontinuous permafrost zone. Non-intrusive geophysical methods offer a possibility to map current permafrost spatial distributions in these environments. In this study, we estimate the depths to the permafrost table ~~surface~~ and base across a peatland in northern Sweden, using ground penetrating radar and electrical resistivity tomography. Seasonal thaw frost tables (at ~0.5 m depth), taliks (2.1 – 6.7 m deep), and the permafrost base (at ~16 m depth) could be detected. Higher occurrences of taliks were discovered at locations with a lower relative height of permafrost landforms indicative of lower ground ice content at these locations. These results highlight the added value of combining geophysical techniques for assessing spatial distributions of permafrost within the rapidly changing sporadic permafrost zone. For example, based on a ~~simple thought experiment~~back-of-the-envelope calculation for the site considered here, we estimated that the thickest permafrost could thaw ~~out~~ completely within the next two centuries. There is a clear need, thus, to benchmark current permafrost distributions and characteristics particularly in under studied regions of the pan-A~~a~~rctic.

1 Introduction

Permafrost peatlands are widespread across the Arctic and cover approximately 12 % of the arctic permafrost zone (Hugelius et al. 2013, Hugelius et al. 2014). They often occur in ~~areas climatically marginal to~~sporadic permafrost areas, protected ~~due~~ by~~to~~ the ~~thermal properties of~~ peat cover, which insulates the ground from heat during the summer (Woo, 2012). In the sporadic permafrost zone the permafrost ground temperature ~~in these peatlands~~ is often close to 0°C, and therefore ~~sensitive to~~even small ~~fluctuations~~ increases in ~~climate~~temperature can result in thawing of permafrost. In addition, p~~P~~ermafrost distribution and thawing in these landscapes is influenced by several factors other than climate, including hydrological, geological, morphological and erosional processes, that often combine in complex interactions (e.g., McKenzie and Voss, 2013; Painter et al., 2013; Zuidhoff, 2002). Due to ~~complex~~ these interactions, ~~between these factors these~~ peatlands are often dynamic with regards to their ~~permafrost~~ thermal structures and extent as the distribution of permafrost landforms (such as dome shaped palsas and flat-topped peat plateaus) and talik landforms (such as hollows, fens and lakes) vary with climatic and local conditions (e.g., Sannel and Kuhry, 2011; Seppälä 2011; Wramner, 1968). This dynamic nature and variable spatial extent has potential implications across the pan-Arctic as these permafrost peatlands store large amounts of soil organic carbon (Hugelius et al., 2014; Tarnocai et al., 2009). The combination of large carbon storage and high potential for thawing make permafrost peatlands biogeochemical hotspots in the warming Arctic. In light of this, predictions of future changes in these environments require knowledge of current permafrost distributions and characteristics, which is sparse in today's scientific literature.

While most observations of permafrost ~~and its condition across the landscape~~ to ~~-~~date consist of temperature measurements from boreholes, advances in geophysical methods provide a good complement for mapping ~~the~~ permafrost distributions in space. Such techniques can provide information about permafrost thickness and the extents and distribution of taliks, which can usually not be obtained from borehole data alone. As the spatial distribution and extent of permafrost directly influences the flow of water through the terrestrial landscape (Sjöberg et al., 2013), adding knowledge about the extent and coverage of permafrost could substantially benefit development of coupled hydrological and carbon transport models~~ing~~ in northern latitudes (e.g., Jantze et al., 2013; Lyon et al., 2010). This may be particularly important for~~to~~ regions where palsa peatlands make up a large portion of the landscape mosaic and regional-scale differences exist in carbon fluxes (G~~e~~iesler et al., 2014).

Geophysical methods offer non-intrusive techniques for measuring physical properties of geological materials; however, useful interpretation of geophysical data requires other types of complementary data, such as sediment cores. Ground Penetrating Radar (GPR) has been used extensively in permafrost studies for identifying the boundaries of permafrost (e.g., Arcone et al., 1998; Doolittle et al., 1992; Hinkel et al., 2001; Moorman et al., 2003), characterizing ground ice structures (De Pascale et al., 2008; Hinkel et al., 2001; Moorman et al., 2003), and estimating seasonal thaw depth and moisture content of the active layer (Gacitua et al., 2012; Westermann et al., 2010). Electrical Resistivity Tomography (ERT) has also been widely applied in permafrost studies (Hauck et al., 2003; Ishikawa et al., 2001; Kneisel et al., 2000), the majority of which focus on mountain permafrost. By combining two or more geophysical methods complementary information can often be acquired raising the confidence in interpretations of permafrost characteristics (De Pascale et al., 2008; Hauck et al., 2004; Schwamborn et al., 2002). For example, De Pascale et al. (2008) used GPR and capacitive-coupled resistivity (CCR) to map ground ice in continuous permafrost and demonstrated the added value of combining radar and electrical resistivity measurements for the quality of interpretation of the data. While some non-intrusive geophysical investigations have been done in palsa peatland regions (Dobinski, 2010; Doolittle et al., 1992; Kneisel et al., 2007; Kneisel et al., 2014; Lewkowicz et al., 2011), the use of multiple geophysical techniques to characterize the extent of permafrost in palsa peatland environments has not been employed.

In this study we use GPR and ERT in concert to map the distribution of permafrost along three transects (160 to 320 m long) in the Tavvavuoma palsa peatland in northern Sweden. Our aim is to understand how depths of the permafrost table~~ surface~~ and base vary in the landscape and, based on resulting estimates of permafrost thickness, to make a first order assessment of the ~~vulnerability~~ potential thaw rate of this permafrost due to climate warming. Further we hope to demonstrate the added value of employing comple~~i~~mentary geophysical techniques in such landscapes. This novel investigation thus helps contribute to our understanding of the current permafrost distribution and characteristics across palsa peatlands creating a baseline for future studies of possible coupled changes in hydrology and permafrost distribution in such areas.

2 Study area

Tavvavuoma is a large palsa peatland complex in northern Sweden at 68°28'N, 20°54'E, 550 masl (Fig. 1) and is a patchwork of palsas, peat plateaus, thermokarst lakes, hummocks and fens. Ground temperatures and weather parameters have been monitored at the site since 2005 (Christiansen et al. 2010). Sannel and Kuhry (2011) have ~~analyzed~~analysed lake changes in the area and detailed local studies of palsa morphology have been conducted by Wramner (1968, 1973).

Tavvavuoma is located on a flat valley bottom, in piedmont terrain with relative elevations of surrounding mountains about 50 m to 150 m above the valley bottom. Unconsolidated sediments, observed from two borehole cores (points 1 and 2 in Fig. 1), are of mainly glaciofluvial and lacustrine origin and composed of mostly sands, loams and coarser grained rounded gravel and pebbles (Ivanova et al., 2011). The mean annual air temperature is -3.5°C (Sannel and Kuhry 2011), and the average winter snow cover in Karesuando, a meteorological station approximately 60 km east of Tavvavuoma, is approximately 50 cm although wind drift generally gives a thinner snow cover in Tavvavuoma (Swedish Meteorological and Hydrological Institute, www.smhi.se/klimatdata/meteorologi).

Permafrost occurs primarily under palsas and peat plateaus in Tavvavuoma, where the average thickness of the active layer is typically 0.5 m (Christiansen et al., 2010; Sannel and Kuhry, 2011). The mean annual temperature in permafrost boreholes (at 2 and 6.1 m depth) in the peatlands of Tavvavuoma range from -0.3°C to -0.4°C (Christiansen et al., 2010). However, no observations of the depth to the permafrost base have been presented for the area. Warming of the air temperature of about 2°C has been observed in direct measurements from the region over the past 200 years (Klingbjer and Moberg, 2003). In light of this warming, winter precipitation (mainly snow) in northern Sweden shows increasing trends over the past 150 years (Alexandersson, 2002). Further, permafrost is degrading across the region and northern Sweden (Sjöberg et al., 2013). For example, peatland active layer thickness in Abisko (located about 60 km south-west of Tavvavuoma) is increasing according to direct observation over the past 30 years (Åkerman and Johansson, 2008) and inference from hydrologic shifts over the past century (Lyon et al., 2009). This regional permafrost degradation has led to changes in palsas as well. Regionally, reductions in both areas covered by palsas and palsa height have been observed (Sollid and Sorbel, 1998; Zuidhoff, 2002; Zuidhoff and Kolstrup, 2000). In Tavvavuoma, both growth and degradation of palsas have been observed in detailed morphological studies during the 1960's and 1970's (Wramner, 1968; Wramner, 1973) and expansion and infilling of thermokarstic lakes have been observed through remote sensing analyses (Sannel and Kuhry, 2011). Palsa degradation and infilling of lakes with fen vegetation have been the dominating processes during recent years (Sannel and Kuhry, 2011; Wramner et al., 2012).

3 Theory and methods

Measurements of permafrost extent and structure were made with both GPR and ERT between 20 August 2012 and 26 August 2012 along three transects covering the main permafrost landforms in the Tavvavuoma area (Fig. 1). The ERT transects were somewhat extended (i.e. slightly longer) compared to the GPR transects to increase the penetration depth along the overlapping parts of the transects. ~~These summer-time measurements were targeted to capture the potential maximum active layer thicknesses in the region. In addition, GPR measurements were also made along the same transects between 22 March 2012 and 24 March 2012 to explore the winter-time conditions. It was not possible to conduct winter-time measurements with ERT due to large snow coverage and frozen surface conditions.~~

Transect T1 was 160 m long and crossed a peat plateau that was raised approximately 1.5 m above the surrounding landscape (Fig. 1). It further crossed two thermokarst depressions (centred at 45 and 130 m) within the peat plateau. Transect T2 was 320 m long, but the southern part covering about 180 m could not be measured with GPR during the summer campaign due to dense vegetation cover (mainly salix sp.). Transect T2 started ~~on~~ on a peat plateau surface at the edge of a drained lake and continued north over a fen (110-180 m) and a small stream (140 m). The northern part, measured with both ERT and GPR, crossed a palsa (200 m) that was raised about 4 m above the surrounding landscape. This palsa has been described via a bore-hole profile (Ivanova et al. 2011, point 1 in Fig. 1). Transect T2 then continued across two fens (250 and 290 m) separated by a lower palsa (270 m). Transect T3 was 275 m long. It started on a relatively low palsa and stretched over a flat area covered by hummocks and thermokarst depressions.

In addition to the geophysical investigations (details of which are described in the following sections), the depth to the permafrost table (the active layer) was probed every 2 meters along all transects using a 1 m steel rod. Sediment cores were retrieved at four points along T1 and two points along T3 down to 2 m. These cores were used to locate the depth to the peat-mineral substrate interface and the depth to the permafrost table (at points 3, 4, 5, and 6~~5~~ in Fig. 1). The topography was measured along the transects using a differential GPS with supplemental inclinometer observations along profiles where only ERT was used~~made using an inclinometer where only the ERT was used~~. The position of the transects was measured using a tape measure and marked at regular intervals to ensure that locations of GPR and ERT transects coincided.

3.1 Ground penetrating radar

Ground penetrating radar (GPR) can be used to map near surface geology and stratigraphy because of differences in dielectric properties between different subsurface layers or structures. An electromagnetic pulse is transmitted through the ground and the return time of the reflected pulse is recorded. The resolution and penetration depth of the radar signal depends on the characteristics of the transmitted pulse and the choice of antennas, which usually range between 10 to 1000 MHz. Higher frequencies will yield a higher resolution but a smaller penetration depth, however, the penetration depth will also depend on dielectric and conductive properties of the ground material. Mapping of permafrost using GPR becomes possible due to the difference in permittivity between unfrozen and frozen water.

In this study, ~~GPR~~ measurements were made with a Malå GeoScience ProEx GPR system using 200 MHz unshielded antennas ~~on~~ along T1 and T2. The transmitting and receiving antennas were held at a constant distance of 0.6 m (common offset) and the sampling time window was set to 621 ns, with recorded traces stacked 16 times~~, with the transmitting and receiving antenna held at a constant distance of 0.6 m (common offset)~~. Measurements were made at every 10 cm along the length of these two transects. Along T3 ~~GPR~~ measurements were made using 100 MHz unshielded antennas with a 1 m antenna separation and measurements made every 0.2 s while moving the antennas along the transect. The sampling time window for T3 was 797 ns and traces were stacked 16 times. The GPR data were processed for a time-zero correction,~~using~~ and with a dewow filter, a vertical gain, and a normal-move out correction for ~~the distance between the transmitting and receiving antennas~~antenna geometry using the software ReflexW (version 6.1, Sandmeier, 2012, ~~downloaded from~~ www.sandmeier-geo.de).

The depths to the permafrost table and the interface between peat-mineral substrates were calculated by converting the two-way travel time to known substrate transitions using estimated velocities for the speed through three different substrate materials: dry peat, saturated peat, and saturated mineral substrate (see Fig. 2 for conceptual sketch of these substrate layers and velocity profiles). To account for uncertainty due to small scale heterogeneity of these ground materials, in addition to the optimal ‘representative’ velocity identified, the likely maximum and minimum velocities for each substrate were considered in the GPR depth conversions (Table 1). The velocity in dry peat (found in the active layer of palsas, hummocks, and peat plateaus) ~~were~~was calibrated ~~from~~ using the active layer thickness measurements made with a steel rod. The minimum and maximum velocities were obtained by subtracting and adding one standard deviation of the measured depths, respectively. For velocities in saturated peat, which was found in taliks such as fens, the depths of saturated peat identified ~~from~~by coring with a 2 m steel pipe (points 3 and 5, Figure 1) were used. The velocity in pure water was used as the minimum velocity and the representative velocity for dry peat was used as the maximum velocity for saturated peat. To obtain ~~these substrate~~ velocities for unfrozen saturated mineral substrate, a common midpoint (CMP) GPR profile~~s~~ was~~ere~~ measured on a drained lake surface (point 7 in Figure 1) ~~at two points. These were made~~ by moving the GPR antennase apart from each other in 10 cm increments along the 15 m long transect~~s~~ (see Appendix A for a detailed description). The CMP approach, thus, allows for ~~recording~~ imaging the same point in space with different antenna offsets making it possible to back out material velocity estimates. ~~The first CMP profile was recorded on a dry peat plateau surface and the second profile on a drained lake surface (points 6 and 7 in Fig. 1). These locations were chosen because they are relatively flat and have flat and uniform reflectors making them suitable for CMP measurements.~~ CMP profiles were analysed using semblance analysis (Neidell and Taner, 1971) in ReflexW software (version 6.1, Sandmeier, 2012, ~~downloaded from~~ www.sandmeier-geo.de).

~~The speed of the GPR signal was estimated through four different substrates (Table 1). These were, namely, the active layer, talik peat, talik mineral, and frozen ground (see Fig. 2 for conceptual sketch of these substrate layers and velocity profiles). As expected, small scale heterogeneity of these ground materials complicates the interpretation of the CMP results. To account for such uncertainty, in addition to the optimal velocity identified, the maximum and~~

~~minimum likely velocities for each substrate were considered in the GPR depth conversions.~~ D~~Further, d~~ue to inherent difficulties with CMP approaches associated with heterogeneous ground substrate properties, ~~unrealistic~~ the velocity estimates from this analysis were complemented with literature defined values to define the ~~for similar substrates~~representative and minimum velocities. The end product here is a range of plausible substrate velocities accounting for potential uncertainties such that any resultant interpretation about subsurface conditions and interface locations can be considered robust.~~.~~

~~The end product here is a range of plausible substrate velocities accounting for potential uncertainties such that any resultant interpretation about subsurface conditions and interface locations can be considered robust. These are expressed as what can be considered a 'representative' velocity bounded by a maximum and a minimum velocity. The active layer velocities were used above the frost table on palsa and peat plateau surfaces that generally consist of relatively dry peat material. For unfrozen peat in fens and under surface depressions where there is a deepening of the permafrost table the talik peat velocities were used. The talik mineral velocities were used for unfrozen mineral sediments, which were assumed to be found only under talik peat. Finally, the velocities for frozen ground were used only when interpreting winter images. Since no local estimates could be made during the winter campaign, literature values for the minimum and maximum velocities were used for interpretations of the GPR data.~~

3.2 Electrical resistivity tomography

Direct-current electrical resistivity measurements are based on a measured potential difference between two electrodes (ΔV) inserted with galvanic coupling to the ground and, similarly, two electrodes where current is injected into the ground (I) with a known geometric factor (k) depending on the arrangement of the electrodes. This gives a value of the apparent resistivity (ρ_a) of the ground sub-surface as

$$\rho_a = k \Delta V / I \tag{1}$$

During a tomographic resistivity survey numerous of these measurements are made in lateral and vertical directions~~both the lateral and vertical~~ (by increasing the electrode spacing). The acquired data is subsequently ~~modeled~~modelled to generate an image of the resistivity distribution under the site. ~~Res2dinv v.3.59.64 by Geotomo software was used for the inversion modeling during this study.~~ Values of resistivity vary substantially with grain size, porosity~~e size~~, water content, ice content, salinity and temperature (e.g. Reynolds, 2011), thus, the resistivity of permafrost also varies to a large degree. This makes electrical resistivity tomography (ERT) techniques useful in detecting the sharp contrast between frozen and unfrozen water content within sediments.

At the Tavvavuoma site, measurements of electrical resistivity were made with a Terrameter LS from ABEM and an electrode spacing of 2 m for the T1 transect and 4 m for the T2 and T3 transects. The Wenner array configuration for the electrodes was used due to its high signal-to-noise ratio and for its accuracy in detecting vertical changes over other common array types (Loke, 2010). For the ~~inversion~~ inverse modeling the smoothness-constrained least-square~~s~~ method was applied (Loke and Barker, 1996). The inversion progress was set to stop on the condition ~~that~~ where the change in root mean squared error from the previous iteration was less than 5% (implying convergence of the inversion). The software Res2dinv (v.3.59.64, Geotomo Software, 2010) was used for the inverse ~~modeling~~modelling during this study.

To assess the quality and reliability of the resistivity modeling for the Tavvavuoma site, the depth of investigation (DOI) method (Oldenburg and Li, 1999) was used. This appraisal-technique uses the difference between two inverted models where the reference resistivity parameter is varied to calculate a normalized DOI-index map. From these values a depth at which the surface data is no longer sensitive to the physical properties of the ground can be interpreted. The method has previously been applied in permafrost studies (e.g., Fortier et al., 2008; Marescot et al., 2003). To calculate the DOI-index we used a symmetrical two-sided difference scheme where 0.1 and 10 times the average apparent resistivity of the resistivity model was considered (respectively) for the initial reference resistivity parameter. Normalized DOI values higher than 0.1 indicate that the model is likely not constrained by the data and should be given little significance in subsequent model interpretation.

To further validate the ERT interpretations, one shorter transect with 0.5 m electrode spacing was conducted over a palsa. This was used to acquire a local resistivity value~~s~~ for the interface between unfrozen and frozen sediments at the bottom of the active layer (surface of the permafrost table). This value (1700 Ωm) allowed us to map permafrost boundaries in the

ERT images, with all resistivity values > 1700 Ωm interpreted as permafrost. However, as the resistivity of the ground varies with other sediment physical properties and the sediment distribution is complex at the site, the resistivity boundary value for permafrost will naturally vary along transects and with depth. For instance, sands generally have maximum values for the unfrozen state close to 1200 Ωm and for some gravels this can reach up to 3000 Ωm (Hoekstra et al., 1974). Finer sediments, such as clays and silts have lower values, ranging from ca 80 to 300 Ωm (Hoekstra et al., 1974). At our site sands dominate, but there is also evidence of loams. Lewkowicz et al. (2011) report a resistivity of 1000 Ωm at the base of permafrost under a palsa in similar, but somewhat finer, sediments conditions in southern Yukon. This value from Lewkowicz et al. (2011) was thus used as a possible minimum resistivity value for the permafrost boundary in the interpretations, while the local resistivity estimate (1700 Ωm) was used as a maximum and representative value. All resistivity values < 1000 Ωm were thus interpreted as unfrozen ground and the values between 1000 and 1700 Ωm represent a range of uncertainty for the location of the interface between frozen and unfrozen sediments. Again, the motivation here was to account for potential uncertainty allowing for robust interpretation.

3.3 Calculations of active layer and thaw rates

To help put the geophysical measurements and their potential implications for this peatland palsa region in context, the thickness of the active layer as well as first order estimate of long-term thaw rates where estimated using a simple equation for 1-D heat flow by conduction, the Stefan equation (as described by Riseborough et al., 2008):

$$Z = \sqrt{\frac{2\lambda I}{Ln}}, \qquad (2)$$

where Z is the thaw depth, λ is thermal conductivity, I is the thawing degree day index (as described by Nelson and Outcalt, 1987), L is the volumetric latent heat of fusion and n is the saturated porosity of the ground substrate. As a talik is by definition unfrozen ground occurring in a permafrost area, the Eq. (2) was used to confirm that ~~the~~ ground identified as talik in Tav~~v~~vavuoma through the GPR and ERT images did not correspond to locations of deeper active layer relative to surrounding positions (i.e. provide a confirmation that these sites would not freeze during winter).

Calculations of active layer depths in fens were made using as input a sinusoidal annual air temperature curve generated from the average temperature of the warmest and the coldest months of the year. The effect of the snow cover, which would give higher ground surface temperatures in the winter, was not explicitly taken into consideration in this simple calculation as we did not have any direct estimates of snow cover available for the transects. As such, these calculations are simply a first-order approximation. Representative properties for saturated peat where chosen, including a thermal conductivity of 0.5 W/m/K and a saturated fraction of 0.80 (Woo, 2012).

In addition, a first-order approximation of long-term thaw rates was carried out ~~out as a thought experiment~~. An instantaneous increase in air temperature of 2° C was assumed, which represents a warming within current climate projections for the 21st century, although at the low end of projections for Arctic warming (IPCC RA5, 2013). A ~~thermal conductivity of 3 W/m/K and a~~ saturated fraction of 0.5~~0~~ ~~were~~ was used to represent ~~a~~ sand ~~soil,~~ slightly oversaturated with ice. To account for some of the uncertainty in this rough estimate, a range of likely minimum and maximum values for thermal conductivity (2 and 3 W/m/K, respectively) for this material, were used to estimate a range of thaw rates. The annual freezing degree days were subtracted from the annual thawing degree days, *I* in Eq. (2), and the amount of days necessary to thaw the estimated local thickness of permafrost was estimated. This is a simple estimate since, clearly, the Stefan equation is not designed to calculate long-term thaw rates nor does such an estimate consider any density dependent feedbacks and/or subsequent hydroclimatic shifts. Regardless, combined with estimates of permafrost thickness made in our geophysical investigation, the aim of this ~~thought experiment~~back-of-the-envelope calculation was to provide an order-of-magnitude estimate for the time it could potentially take permafrost to completely thaw ~~out~~ at this site to help place it in a pan-arctic context.

4 Results

4.1 GPR data

In the ~~summer~~ GPR images the permafrost table was clearly detectable under the palsa and peat plateau surfaces along all transects (Fig. 3). The interface between peat and mineral substrates was only detectable in unfrozen sediments. Deeper reflections, interpreted as the permafrost table under supra-permafrost taliks, were found under the fens and surface depression in all transects. At~~In~~ the beginning of both transects T1 and T2, deep reflections that end abruptly were present in the images at about 250 ns and 150 ns, respectively. In T1, this corresponds to a wet fen bordering a lake and for T2 it corresponds to a fen bordering a stream. The proximity to these water bodies suggests that these are likely not reflections from the permafrost table. The base of the permafrost could not be detected at any point in the GPR images likely because of loss of signal strength at depth.

~~In the winter GPR images, interfaces between frozen and unfrozen ground (taliks) were not clearly separable from sedimentary layers. Several interfaces are visible in the images, however, indicating a complex layering of sediments and ground ice confirmed by described sedimentary profiles from the site (Ivanova et al., 2011). No further interpretations were made from winter images, however, due to the lack of clear reflections.~~

4.2 ERT data

The ~~results from the ERT data modelling~~inverted resistivity sections showed areas of high resistivity (1000-100000 Ωm) where permafrost could be expected due to the sharp contrast to surrounding surfaces. This suggests permafrost boundaries are detectable for both the extent of the horizontal distribution and the vertical extent to the base of permafrost (Fig. 4). The highest resistivity values were found under the peat plateau in T1 and under the palsas in T2 and T3. Low resistivity values were found under the fens in all transects. DOI values increase with depth for all transects allowing the permafrost base to be interpreted only along parts of T2. ~~Counter to this~~In contrast, under T1 and T3 the DOI rapidly increases under the peat plateau and hummocks. Due to the wide electrode spacing adopted (2 and 4 m), the permafrost table under the active layer is too shallow to be visible in the ERT data.

4.3 Geophysical interpretations

Permafrost occurs under the palsa and peat plateau surfaces along T1 and T2, as well as under the hummocks along T3 (Fig. 5). The active layer depths estimated from the GPR data closely matched the depths measured in the field (Table 2). This is expected since measured active layer depths~~ along T1 and T2~~ were used to derive the velocity of the radar signal in the dry peat in the active layer. The depth to the base of the permafrost could only be estimated with good confidence along parts of T2 and is on average 15.8 m from the ground surface and at least 25 m at its deepest point. Along transects T1 and T3 the deepest permafrost was found at 8.4 m and 23.4 m respectively; however, the permafrost base could not be identified with confidence below this depth.

Potential taliks (Table 3 and Fig. 5) are numerous and occur in both wet fens, such as all taliks along T2, and relatively dry depressions in the terrain, such as all taliks along T1. The sediment cores used for estimating the GPR representative signal velocity in saturated ~~talik~~ peat were taken in both a relatively dry location and in a wet fen, but the calculated velocities were nearly identical, indicating that the soil moisture at depth was similar at both locations. Most of T3 was underlain by taliks and these were found under both wet fens and drier surface depressions. The taliks range in depth from 2.1 m (T3f, numbering from Table 3 and Fig. 5) to 6.7 m (T1c) based on the GPR data and are slightly deeper, however within the range of uncertainty, based on the ERT results. From the ERT data, T1c is in fact interpreted as a potential through-going talik. Talik T1b was only detected from the ERT data, and taliks T3b – T3d appear as one large talik in the ERT data.

4.4 Calculations of active layer and thaw rates

The active layer depths calculated using the Stefan equation support the interpretation that identified taliks do not freeze during winter. The seasonal frost penetration depth was estimated to be 0.72 m which is about the same as the average peat depth along the transects and much less than the estimated minimum depth of the taliks (2.1 m). While a shallower peat depth would give a deeper frost penetration it is unlikely that the seasonal frost penetration is > 2.1 m in the area surrounding Tavvavuoma. This ancillary estimate confirms the aforementioned geophysical interpretation. Further, ~~as a thought experiment,~~ assuming a 2°C instantaneous temperature increase at the site, ~~the~~ a first order approximation of the long-term thaw rate was calculated to be ~~7~~ 6 - 8.5 cm/year. At this rate, the time to completely thaw ~~out~~ permafrost assuming the estimated average thickness along T2 (15.3 m) was calculated to be 175 - 260 years.

5 Discussion

5.1 Permafrost and talik distribution at Tavvavuoma

The spatial pattern of permafrost and taliks in Tavvavuoma is closely linked to the distribution of palsas, peat plateaus, fens and water bodies. This suggests that local factors, such as soil moisture, groundwater flow, ground ice content, sediment distributions and geomorphology, strongly influence the local ground thermal regime (see e.g. Delisle and Allard, 2003; McKenzie and Voss, 2013; Woo, 2012; Zuidhoff, 2002). The relative elevation of permafrost landforms, as well as permafrost resistivity values and sediment distributions suggest that there is a large variation in ground ice content in the area. Surface elevations of palsas and peat plateaus are highest along T2 and lowest along T3, indicating a higher ice content of the underlying ground along T2, which is likely related to differences in ground substrates between the transects. Coring (<2 m) across the site, as well as existing borehole descriptions (Ivanova et al., 2011) confirm that the ground contains a larger fraction of coarse glaciofluvial sand and gravel, which are not susceptible to frost heave, closer to T3 as compared to T2.

Lewkowicz et al. (2011) used the height of palsas and permafrost thickness, estimated by ERT, to calculate excess ice fractions (EIF =defined as the ratio of –palsa height to ~~/~~permafrost thickness) in permafrost mounds in southern Yukon. In~~For~~ Tavvavuoma,~~ the EIF at the top of~~ the highest palsa at T2 is approximately 4 m high and underlain by 16 m thick permafrost at the highest point. This corresponds to an EIF of 0.25~~was 0.25,~~ which is comparable to the ~~those~~ EIFs reported by Lewkowicz et al. (2011), which were generally ranging between 0.2 and 0.4. In contrast, along T3 the relative heights of permafrost landforms are lower and the permafrost is thicker for most of the transect. Similarly calculated EIFs along T3 were on average <0.03 and at maximum <0.09, but are likely lower in reality as the base of the permafrost is at a greater depth than what could be detected in our study. ~~For T3 possible maximum EIFs were calculated using the greatest depths at which permafrost was found, since the permafrost base could not be identified from the data for T3. The calculated EIFs along T3 were on average <0.03 and at maximum <0.09, confirming that ground ice content is lower along this transect.~~ The relatively low resistivity of the permafrost along T3 further supports interpretations for lower ice content in this permafrost. Permafrost with low ice content is more susceptible to thaw, as less energy is needed for latent heat exchange. This provides a possible explanation for why taliks are more widespread along T3, as permafrost with a low ice content would have reacted more rapidly to warming in the area.

The calculated thaw rate of 6 – 8.5~~7~~ cm/year is considerably higher than the circa 1 cm/year deepening of the active layer observed in the region (Åkerman and Johansson, 2008) and inferred from hydrological records (Lyon et al. 2009). One possible reason for this is that these observations were made in the relatively ice rich top layer of peat, while for the calculations in this study a medium with higher thermal conductivity and lower ice content was used to represent the lower mineral sediment layer. The 2°C instantaneous step change in temperature could further have contributed to the higher thaw rates compared to the ones observed. As thawing is driven by gradients in heat it can be argued that permafrost thaw rates should increase with warmer air temperatures. Considering this, the calculated time of complete permafrost thaw ~~out~~ of about 175 – 260 years can be considered reasonable in at least an order of magnitude. However, much more rapid palsa degradation has been observed in the region (Zuidhoff, 2002), due to block and wind erosion processes and thermal influence on palsas from expanding water bodies, and very rapid decay of palsa surface areas has been observed in both southern Norway and the Canadian Arctic (Payette et al., 2004; Sollid and Sorbel, 1998). The coupled erosion, hydrological and thermal processes are not represented in the Stefan equation but can be of great importance for permafrost thaw rates (McKenzie and Voss, 2013; Painter et al., 2013; Zuidhoff, 2002). There is clearly a need for quantification of the relative importance of these processes for permafrost thaw to better understand expected future changes in these environments.

5.2 On the complementary nature of the geophysical techniques

Several previous studies have shown the benefits of combining more than one geophysical technique for mapping permafrost (e.g. De Pascale et al., 2008; Hauck et al., 2004; Schwamborn et al., 2002), and also in this study the GPR and ERT data provided complementary information that allowed for interpretations that would not have been possible by using only one of the two datasets. Of course, combining multiple techniques for inference compounds our estimate uncertainties. To attain more precise estimates of depths to the different interfaces, deeper coring data would have been necessary for both more accurate signal velocity estimates for the GPR and for local resistivity values of the ground materials. The fact that ERT depth estimates are consistently higher than the GPR estimates suggest that either the resistivity boundary value for permafrost is in fact lower than our local estimate, or that GPR signal velocities are higher than the values used in this study. Since our local permafrost resistivity estimate was made in peat at the permafrost table, which can have a very high ice content compared to deeper sediment layers, it is a more likely explanation for this discrepancy.

GPR and ERT yielded somewhat overlapping data but the two datasets have different strengths and therefore complement each other well. The GPR data worked well for identifying the permafrost table with high confidence, especially in the top 2 meters where sediment cores could be easily obtained for validation and signal velocity estimates. This suitability of GPR for identifying permafrost interfaces in the top 1-2 meters has been shown in several studies (e.g. Doolittle et al., 1992; Hinkel et al., 2001; Moorman et al., 2003). ~~GPR was however not a suitable technique for detecting the permafrost base during winter, due to the high variability in ground ice content and sediment layering in this environment.~~ The ERT data, using the setup in this study, does not yield data in the uppermost part of the ground and also has higher uncertainty where resistivity contrasts are high (Fig. 4), which makes it less well suited for the active layer and shallow taliks. With the ERT data it is, however, possible to image relatively deep in the ground, where the GPR cannot penetrate. By combining both GPR and ERT the active layer, the base of permafrost, and potential taliks could be identified along at least parts of the transects, which could not have been achieved with good confidence by either of the two methods alone.

6 Concluding remarks

Peat plateau complexes offer an interesting challenge to the Cryosphere community as they are clear mosaics combining local-scale differences manifested as permafrost variations. As such variation occurs both horizontally and vertically in the landscape, geophysical techniques offer a good possibility to record current permafrost conditions across scales. ~~Complementary~~ Further, by combining methods, such as GPR and ERT as demonstrated here, ~~are necessary as they can provide orthogonal~~ complementary and independent views of the permafrost extents can be acquired. The results of this study show a heterogeneous pattern of permafrost extent reflecting both local and climatic processes of permafrost formation and degradation. To improve our understanding of landscape-permafrost interactions and dynamics will require a community effort to benchmark variability across the scales and environments within the pan-A~~a~~rctic. This is particularly important in lesser studied regions and across the sporadic permafrost zone where changes are occurring rapidly.

Appendix A: CMP analyses

Common mid-point (CMP) analysis is a widely used method to estimate local GPR signal velocities through ground materials. By moving GPR transmitting and receiving antennas apart incrementally between measurements, the same point in space is imagined with different antenna offsets making it possible to back out material velocity estimates from the hyperbolic shape of the recorded reflectors. The measured reflectors must be relatively flat so that the signal moves through the same materials at the same depths independent of antenna offset.

In this study, a CMP profile was measured on a drained lake surface (point 7 in Figure 1). Coring down to 2 m with a steel pipe at this location revealed the existence of an unfrozen saturated peat layer down to 1.75 m depth and unfrozen mineral soils consisting of mainly sand and silt below that depth. The aim of the CMP measurement used in this study was to estimate material velocities for the deeper unfrozen saturated mineral substrate. As such, deeper here means materials below the observation depths of the permafrost probe (< 1 m) and coring with the steel pipe (< 2 m). For the CMP measurement, 100 MHz unshielded antennas were moved apart in 10 cm increments along a 15 m transect with a time window of 797 ns and 16 stacks of each trace.

The measured CMP data was analysed in ReflexW (version 6.1, Sandmeier, 2012, downloaded from www.sandmeier-geo.de). The data were processed for a time-zero correction, a dewow filter, and a vertical gain. Semblance analysis (Neidell and Tanner, 1971) was used to identify appropriate reflectors from which velocities could be estimated. Using semblance analysis numerous velocities are used to correct the recorded reflectors for the increased travel distance for each offset. When the hyperbolic reflection becomes coherent in the CMP (i.e. a flat line) a high semblance value is obtained indicating that the correct velocity has been used.

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Figure A1shows the estimated velocity profile, recorded CMP radargram, and semblance plot for the CMP transect. Although a relatively flat reflector was identified for the CMP measurement, the results from the semblance analysis does not show one clear reflector and associated velocity at the identified depth of the peat-mineral interface. Instead, a wide range of possible velocities are shown in the semblance plot for the top ~200 ns, likely due to high heterogeneity in ground substrates and/or water content. Due to the difficulty in constraining the material velocities for the deeper layers using this method, these results were only used for estimating a probable maximum velocity in unfrozen mineral sediments (as this was higher than most literature values). This maximum velocity estimate was complemented with literature values for the representative and minimum velocities.

Author contribution

Ylva Sjöberg designed the study, carried out the GPR measurements and analysis, and did the main writing of the manuscript. Per Marklund carried out the ERT measurements and analysis and did the main writing for the sections on ERT methods and ERT results, as well as commented on the whole manuscript. Rickard Pettersson provided input on the geophysical techniques and analyses and commented on the whole manuscript. Steve Lyon provided input on the project design and commented on the whole manuscript, including language and style.

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Table 1. Velocities used for converting two-way travel times to depth in GPR data.

Material	Velocity (m/ns)	Method/Source
~~Active layer~~Dry peat – representative	0.049	Calibration against every second field measurements[i] of active layer depths
~~Active layer~~Dry peat - min	0.046	Representative estimate minus one standard deviation of field measurements[i]
~~Active layer~~Dry peat - max	0.052	Representative estimate plus one standard deviation of field measurements[i]
~~Talik~~ Saturated peat – representative	0.036	Calibration against coring (point 3 and 5, in Fig. 1)
Saturated~~Talik~~ peat - min	0.033	Velocity in pure water (Davis and Annan, 1989)
Saturated~~Talik~~ peat - max	0.049	Representative estimate for ~~active layer~~dry peat
Saturated~~Talik~~ mineral – representative	0.060	Velocity in sand and clay from Davis and Annan (1989)
Saturated~~Talik~~ mineral - min	0.053	Calculated from Joseph et al. (2010) for saturated loams and sands
Saturated~~Talik~~ mineral - max	0.073	Highest estimated velocity from CMP analysis
~~Frozen ground – min~~	~~0.110~~	~~Minimum for permafrost from Hinkel et al. (2001)~~
~~Frozen ground – max~~	~~0.160~~	~~Velocity in pure ice from Evans (1965)~~

[i] Field measurement using a 1 m steel rod

Table 2. Range of interpreted depths (m) of active layer, peat-mineral interface, and permafrost base averaged along transects at Tavvavuoma.

	T1			**T2**			**T3**		
	Min[i]	**Representative**[ii]	Max[iii]	Min[i]	**Representative**[ii]	Max[iii]	Min[i]	**Representative**[ii]	Max[iii]
Active layer									
Observed[iv]		**0.51**			**0.52**			**0.56**	
GPR	0.50	**0.53**	0.57	0.48	**0.51**	0.54	0.52	**0.56**	0.59
Peat-mineral interface									
GPR	0.77	**0.84**	1.14	0.68	**0.74**	1.01	0.63	**0.69**	0.93
Permafrost base									
ERT		**-**	-		**15.8**	17.3		**-**	-

[i] GPR: using the estimated minimum velocity (Table1). ERT: using 1000 Ωm resistivity boundary (talik).

[ii] GPR: using representative estimate velocity (Table 1). ERT: using 1700 Ωm resistivity value.

[iii] GPR: using the estimated maximum velocity (Table 1). ERT using 1000 Ωm resistivity boundary (permafrost base).

[iv] Depth from manual field measurement using a steel probe.

Table 3. Estimated depths (m) of taliks. Numbering is the same as in Fig. 5.

Talik max depth	GPR min[i]	**GPR representative[ii]**	GPR max[iii]	ERT min[i]	**ERT representative[ii]**
T1a	2.4	**2.7**	3.4	2.5	**3.1**
T1b	-	**-**	-	1.6	**2.8**
T1c	6.0	**6.7**	8.3	> 4.7	> **4.7**
T2a	5.4	**6.1**	7.6	5.4	**6.9**
T2b	5.3	**6.0**	7.4	6.9	**8.8**
T3a	5.3	**5.9**	7.4	5.8	**7.8**
T3b	5.7	**6.4**	8.0	6.3	**8.2**
T3c	5.1	**5.7**	7.0	4.8	**7.9**
T3d	3.1	**3.5**	4.4	-	**4.0**
T3e	4.6	**5.2**	6.4	5.4	**7.2**
T3f	2.0	**2.1**	2.2	-	**3.8**
T3g	3.7	**4.1**	5.2	5.0	**6.8**

[i] GPR: using the estimated minimum velocity (table1). ERT: using 1000 Ωm resistivity boundary (talik).

[ii] GPR: using representative estimate velocity (table 1). ERT: using 1700 Ωm resistivity value.

[iii] GPR: using the estimated maximum velocity (table 1). ERT using 1000 Ωm resistivity boundary (permafrost base).

Figure 1. Location of the study site (inset), investigated transects, existing boreholes (Ivanova et al., 2011, points 1 and 2), coring points, and points of CMP measurement~~s~~ (described in section 3.1 and Appendix A). (Aerial photograph from Lantmäteriet, the Swedish land survey, 2012.)

Figure 2. Conceptual sketch of typical distribution of ground substrates and associated estimated velocities for a palsa and talik ground profile.

Figure 3. Elevation profiles and GPR images for T1, T2, and T3 with selected reflections marked as examples of interfaces that were identified for this study. Landforms are indicated on top of elevation profiles along T1 and T2 (Tk.D = thermokarst depression) together with coring points in T1 (a = point 3 in Figure 1, and b = point 4 in Figure 1) and T3 (d = point 6 in Figure 1) as well as the 10 m borehole in T2 (c = point 1 in Figure 1). No landforms are indicated along T3 after the first palsa (0-25 m) due to the complex micro topography of hummocks and thermokarst depressions along this transect.

Figure 4. Elevation profiles and ERT results for T1, T2, and T3. DOI <~~>~~ 0.1 (black lines) indicates that the model is well constrained by the data. Landforms are indicated on top of elevation profiles along T1 and T2 (Tk.D = thermokarst depression) together with coring points in T1 (a = point 3 in Figure 1, and b = point 4 in Figure 1) and T3 (d = point 6 in Figure 1) as well as the 10 m borehole in T2 (c = point 1 in Figure 1). No landforms are indicated along T3 after the first palsa (0-25 m) due to the complex micro topography of hummocks and thermokarst depressions along this transect.

Figure 5. Interpreted permafrost distribution along T1, T2, and T3. Uncertainty intervals come from the range of estimated signal velocities for GPR (Table 1) and from the range of resistivity values (1000-1700 Ωm) used for identifying the permafrost boundary for ERT. In sections marked *GPR Talik* (red dotted line) GPR depth conversions have been made using

saturated peat velocities down to the peat-mineral interface (green line) and then using saturated mineral substrate velocities down to the permafrost table (blue line). In the remaining parts of transects the dry peat velocities have been used down to the permafrost table. No interpretations of ERT data with DOI > 0.1 are have been made and therefore the permafrost base is only visible along parts of T2. Note the differences in scale in the x-direction between figures and the vertical exaggeration.

Figure A1. Estimated velocity profile, recorded CMP radargram, and semblance plot for the CMP transect measured on the drained lake surface. The semblance plot shows more likely velocities in darker shades of grey with the velocities from the reflectors (red lines in radargram) used for generating the velocity profile indicated by black and red diamonds.