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Quantifying Snow and Vegetation Interactions in the High Arctic Based on Ground Penetrating Radar (GPR)

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Abstract

The quantification of the relationship between accumulation of snow and vegetation is crucial for understanding the influence of vegetation dynamics. We here present an analysis of the thickness of the snow and hydrological availability in relation to the seven main vegetation types in the High Arctic in Northeast Greenland. We used ground penetrating radar (GPR) for snow thickness measurements across the Zackenberg valley. Measurements were integrated to the physical conditions that support the vegetation distribution. Descriptive statistics and correlations of the distribution of each vegetation type to snow thickness, as well as to external factors that influence the redistribution of snow were performed. We found that although there is wide variability in the snow packing, there is strong correlation between snow thickness and the distribution of certain plant communities in the area. The accumulation of snow and occurrence of vegetation types such as *Dryas octopetala* heath and *Salix arctica* snowbed showed more influence by the microtopography than by other vegetation types that showed independence of the terrain conditions.

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Introduction

Snow depth and snow cover distribution are important influences on the ecology of local ecosystems since snow is a key factor in the hydrology dynamics of the soil (Jones, 1999). The distribution of snow is mainly determined by the microtopography, wind direction, and the amount of precipitation. The snowpack protects the plants from abrasion, and deeper snow layers reduce desiccation over winter by reducing temperature gradients in snow and soil (Pomeroy and Brun, 2001). The snowpack is eroded from sites with sparse or low vegetation, as well as those that are exposed, and transported to locations with denser, taller vegetation and topographic depressions (Li and Pomeroy, 1997). An additional environmental factor related to snow accumulation is surface albedo, which is a major link to global climate change. Projected decrease in snow cover will not just influence albedo and hence the energy balance, but also the exchange of greenhouse gases between polar landscapes and the atmosphere (Fischlin et al., 2007). Hence, Arctic ecosystems have long been considered among the most sensitive and most influential with respect to climatic changes (Oechel et al., 1993, 1998; Maxwell, 1997).

Seasonal snow cover influences greenhouse gas emissions from the soil through the duration and depth of snow during the cold, dry accumulation period, as well as the discharge rates and chemical composition of meltwaters during the melt season (Van Bochove et al., 1996). Vegetated areas reduce surface wind speeds limiting sublimation events, whereas the local melting processes during spring affect the timing and variability of snowmelt runoff and therefore the length of the growing season (McFadden et al., 2001). In addition to these physical links to vegetation, soil conditions and nutrient availability are also key factors in plant distribution and productivity, which are in turn controlled by climatic conditions and snow-vegetation interactions (Jones, 1999). In some cases, animal activity also has an effect on vegetation and on the redistribution of nutrients through the excretion of feces and urine (Walker et al., 1999). On the other hand, there is also an influence in the variability of overall plant production on the dynamics of foraging animals (Forchhammer et al., 2005).

Precise small-scale mapping of snow accumulation patterns are difficult and time consuming in the remote Arctic. Several geophysical methods have been developed over the years to facilitate efficient mapping of snowpack characteristics over large areas. Among the methods used to quantify snow accumulation and snow density, the most suitable for collecting information quickly and continuously with minimal disturbance is ground penetrating radar (GPR) (Annan et al., 1994; Jaedicke and Sandvik, 2002). GPR has been successfully used in multiple snow-ice studies and has shown high accuracy in the determination of snow cover thickness (Harper and Bradford, 2003; Arcone, 2009). The relatively homogeneous condition of the snowpack in cold regions allows GPR to measure snow thickness and stratigraphy covering large areas, and it provides much progress in glaciological studies with respect to mass balance and snow cover in general (e.g., Gogineni et al., 2000; Arcone et al., 2005; Godio et al., 2008; Heilig et al., 2009).

The ecology of the Zackenberg area in Northeast Greenland has been intensively studied (Meltofte et al., 2008). Tamstorf et al. (2008) presented an analysis of the relationship between snow thickness and classified vegetation cover (greenness) using Normalized Difference Vegetation Index (NDVI). Our study is a contribution to the current knowledge of the snow and vegetation interactions that take part at Zackenberg. This paper aims to quickly and efficiently study small-scale effects of snow accumulation on distribution of vegetation in a region of the High Arctic. GPR data collected was used to analyze and quantify the distribution patterns of snow accumulation in relation to vegetation. The analyses and results here presented are in consideration of factors that contribute to the dynamics of the snow packing, redistribution, and melting processes.

Methodology

SITE DESCRIPTION

The Zackenberg valley is located in Northeast Greenland (74°28′N, 20°31′W; Fig. 1). The valley is part of a major valley

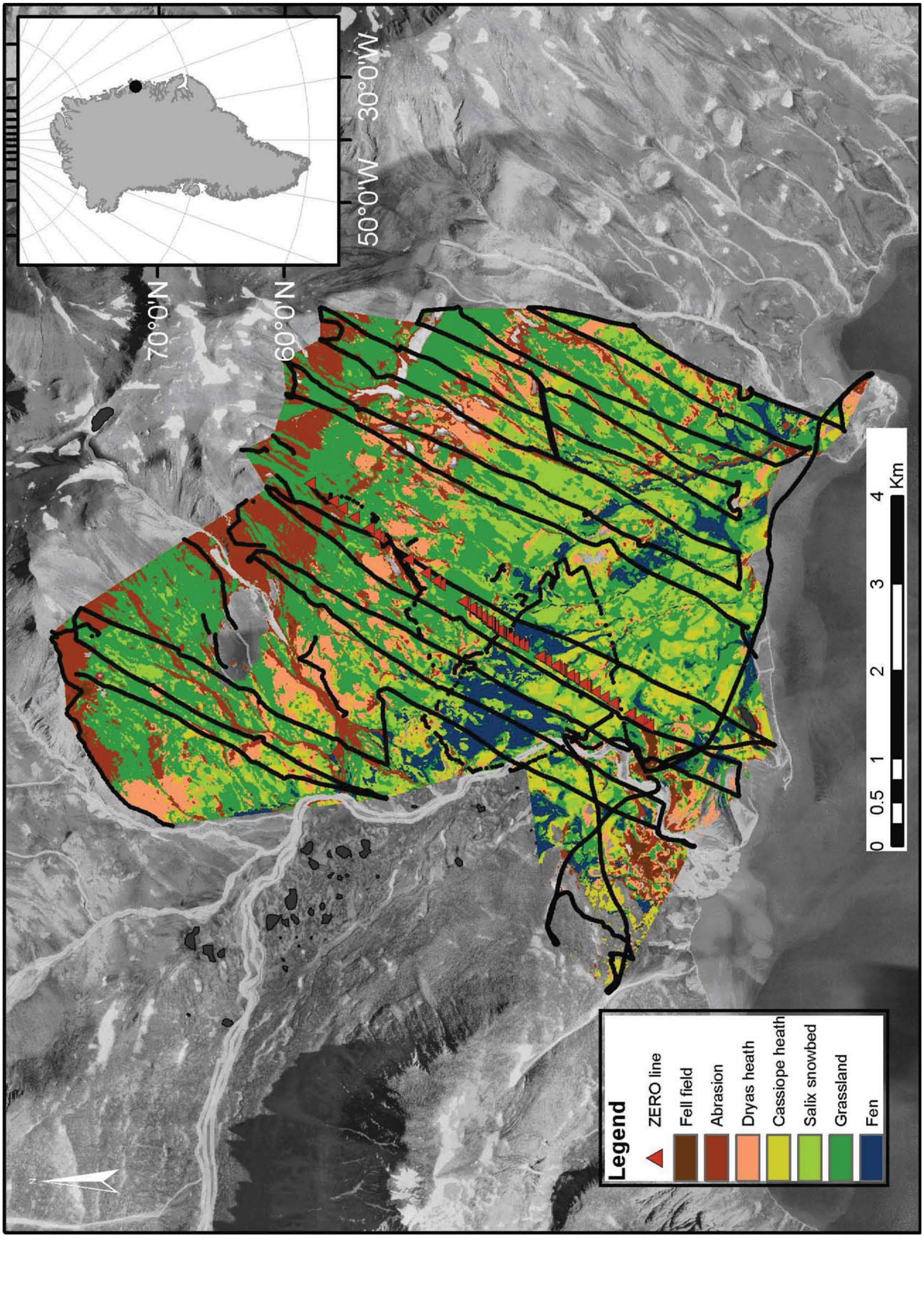

FIGURE 1. Location of Zackenberg area and vegetation map of the studied site. Black lines are measurements of snow thickness using GPR and Global Positioning System (GPS) snow-depth probe. Colors cover the areas that correspond to vegetation types of interest.

system surrounded by mountains and underlain by continuous permafrost. The study area consists of two major landscape units: the lower valley, and the southwest-facing slopes of Aucellabjerg Mountain. The valley floor is generally flat with some slopes along the rivers and streams. In contrast, the slope increases slightly with the elevation going up Aucellabjerg. Above 300 m a.s.l. vegetation becomes sparser. In general, precipitation and high wind speeds in Northeast Greenland in the winter are usually connected to cyclonic activity over the Greenland Sea. Hansen et al. (2008) thoroughly described weather including precipitation for the Zackenberg area, where the prevailing winter wind direction is from north to northwest with small variations; consequently, the larger snow drifts are mainly located on the south-facing slopes (Hinkler et al., 2008).

In the Zackenberg valley, the snow cover generally starts to form in September or early October and usually begins to melt by mid- to late May. The growing season normally starts by the middle of June and ends in late August or early September (Tamstorf et al., 2007). The distribution and physical characteristics of the accumulated snowpack have an influence on the heat flux and thus the chemical composition of the soil, which in turn strongly influences the vegetation mosaic. Additionally variations in the soil texture affect the active layer thickness, small-scale differences in microrelief, soil moisture, and the openness of the plant canopy. In general, in frozen grounds other microscale factors such as ice wedges and frost boils can also cause major differences in the soil thaw depth (Walker et al., 2003).

The distribution of the major plant communities are closely related to these landscape units, the snow cover, soil conditions (e.g., type and moisture content), and elevation. The flora of the area includes more than 150 species (Bay, 1998). The vegetation has been classified into seven plant communities. These communities are characterized by the composition and cover of vascular plant species, the cover of mosses and lichens, and their relation to terrain and soil parameters (Bay, 1998). The land cover map made in 2005 using hyper-spectral imagery with 5 m spatial resolution depicts 14 land cover classes that include the following vegetation types: fen, grassland, two types of dwarf shrub heath (*Cassiope tetragona* heath and *Dryas octopetala* heath), *Salix arctica* snowbed, abrasion plateau, and fell field (Elberling et al., 2008). The map also depicts other land cover classes presented as non-vegetated areas or impediments for mapping: barrens, boulder field, patchy boulder field, river-bed, shadow, water, and snow.

DATA AND MEASUREMENTS

In late March 2008, approximately 48 km^2 of the Zackenberg valley was surveyed using GPR to measure the thickness of the snowpack accumulated during the winter 2007/2008. Since no major precipitation events occurred between late March and the beginning of the melt in May, the amount of snow was similar to that found at the end of the winter snow period.

In this study snow depth was obtained from data collected using an X3M system (Malå Geosciences) with a 500 MHz shielded antenna. The antenna was moved over the surface using a snowmobile and a GPS for precise positioning. Data were collected in random stratified transects across the valley and up the Aucellabjerg slopes (eastern side of the valley), a range of different types of terrain (Fig. 1), with a scan rate collection of 0.19 m/scan. The total transect length measured using the GPR was approximately 124 km. Measurements of snow thickness were made up to a maximum elevation of about 600 m a.s.l. A 3 m steel rod was used for manually probing the snow at regular intervals along the radar transects. The snowmobile was stopped and the trace number registered while measuring the depth manually next to the antenna (15 cm from the center of the antenna). Snow depths above 3 m were noted as >3 m and here no accuracy assessment was possible. Depending on the snow condition (loose/hard snowpack), it was possible to estimate the actual snow depth on the steel rod with an accuracy of ± 1 cm. The normally distributed sample of direct measurements of snow depth is plotted against the estimated values using the GPR method in Figure 2. Confidence intervals of 90% are included for the linear regression ($R^2 = 0.8$, $p < 0.5$) that validates our assumptions of the use of GPR data for this purpose.

The GPR data set was reduced to points separated with a minimal distance of 7.1 m to conform to the spatial resolution (cell size) of the maps used in the analysis. At some specific sites where the snowmobile might have disturbed the vegetation or soil due to a shallow or non-existent snow layer, a GPS snow depth probe (Magnaprobe) was used to measure and register the snow depth and coordinates of each measurement point. Additionally, manual measurements of the snow density were taken over an area of the study site that is representative of the study site (Fig. 3).

GPR is an Ultra High Frequency system that transmits an electromagnetic signal which is then reflected by the electrical heterogeneities between the materials in which it propagates and returned back to the antenna. In this case, the interface between the bottom of the snowpack and the ground appears as a clear reflector in the signal received. The received signal is a measure of the travel time (t) of the electromagnetic waves; as dry snow can be considered a non-conductive, non-dispersive, and non-magnetic

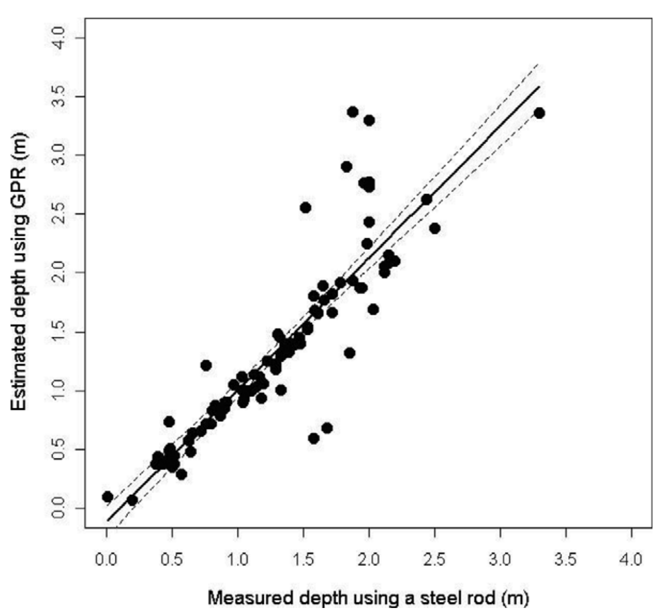

FIGURE 2. Ninety percent confidence intervals (dashed lines) of the linear regression (solid line) between direct measurements of snow depth and estimated values of depth obtained from GPR data.

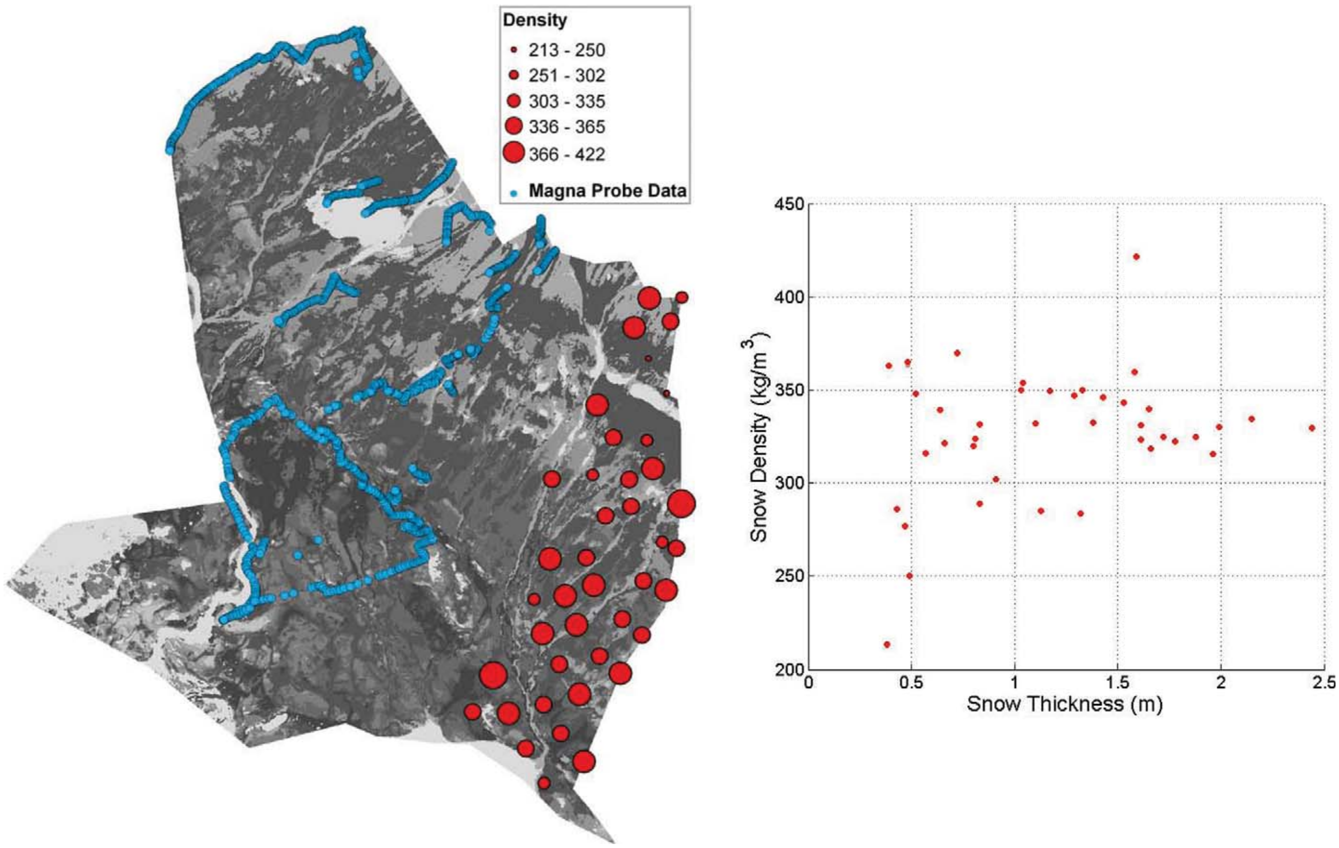

FIGURE 3. Location of manual measurements in the study area. (Left) Snow depth using Magna probe (blue dots) and snow density measurements (red spots). (Right) The scatter plot of depth versus density.

media, velocity of the signal can then be simply determined either at the calibration points where depth (d) is known as $v = 2d/t$ or from the relative dielectric permittivity (ϵ_r) by using

$$v = c / \sqrt{\varepsilon_r} \quad (1)$$

where c is the velocity of electromagnetic waves in a vacuum (3×10^8 m/s).

Variations in the propagation velocity are dependent of the dielectric permittivity. In the snowpack, differences are produced due to changes in the density (porosity, morphology, and metamorphism of the snow crystals), water content, and temperature (Evans, 1965; Bogorodsky, 1985). Snow density ranges between 240 (fresh snow) and 640 kg/m^3 (granular snow). In Zackenberg, densities are periodically measured manually; previous observations show that at the time GPR was used (mid-March), variations were not substantial (Hinkler et al., 2008). Snow density averaged 330 kg/m^3 with a standard deviation (SD) of 0.04 kg/m^3 for the points shown in Figure 3.

The dielectric behavior of dry snow with respect to density can be obtained from the relationship formulated by Looyenga (1965):

$$\varepsilon_{r,snow}^{1/3} - 1 = x\,(\varepsilon_{r,ice}^{1/3} - 1)\; dB / m \quad (2)$$

where x is the ratio between densities of snow mixture and ice (917 kg/m^3). Given that the relative dielectric permittivity of ice is 3.2 (Evans, 1965), snowpack thicknesses were interpreted from the radar data using the estimated wave velocity of 0.235 m/ns in average for this case.

The identification of plant communities in the following analyses was based on a raster map with 14 land cover classes (Elberling et al., 2008). As the vegetation is influenced not only by the thickness of the snowpack but also by the topography represented by slope and aspect of the terrain, elevations above sea level, and wind direction, these factors were added as variables to the analyses. A digital elevation model (DEM) with 5 m resolution was also used in the study area. As wind is predominantly from the north-northwest during the winter period (Hansen et al., 2008), the effect of changing slope-face directions was also analyzed.

Time-domain reflectometer (TDR) measurements as GPR are based on the determination of the travel time of a propagated electromagnetic signal. Because of the homogeneity that accumulated snow presents, GPR efficiently provides continuous information of the subsurface when prospecting large areas for estimating depth or snow layering (e.g. Spikes et al., 2004; Arcone, 2009). In contrast, TDR methods are used for point measurements, usually as a calibration method (Huisman et al., 2000; Previati et al., 2011). Here, estimations of water content from GPR data are compared with TDR measurements that were taken at 35 specific sites at incremental depths of either 5 or 10 cm down to a maximum depth of 40 cm. The point measurements conform to the ''ZERO line'' in Figure 1. This line is part of an ongoing monitoring program and has been previously studied with each point characterized by vegetation (Fredskild and Mogensen, 1997; Hansen et al., 2011).

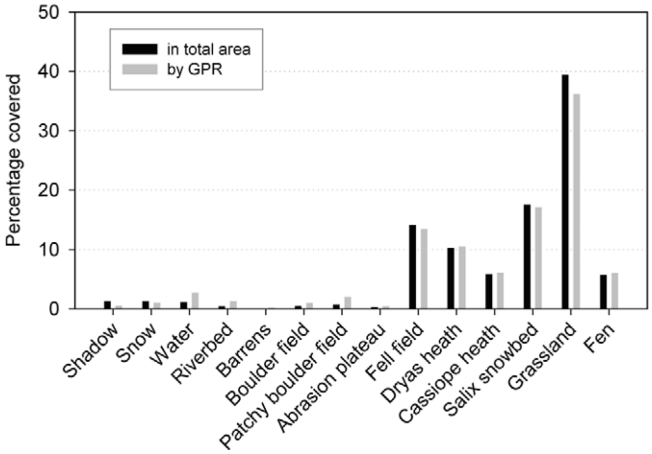

FIGURE 4. Percentage of vegetation and other land-cover classes of the total area of the valley and the area where GPR data were collected.

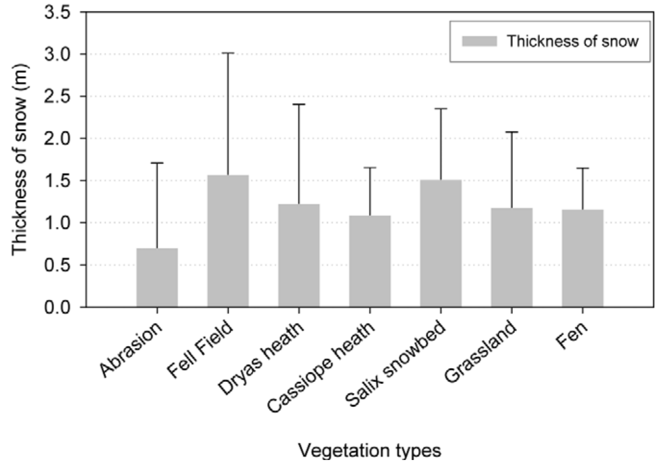

FIGURE 5. Average thickness of the snow for the seven vegetation types studied. Error bars represent the standard deviation.

STATISTICAL ANALYSIS

Assuming the strong hydrological influence that snow has on vegetation, data were separated into the seven vegetation types mentioned previously. From this, we aimed to estimate the dependence of vegetation on snow thickness by quantifying the relation of both parameters (which are in turn influenced by topographical and geological situation) using correlation analysis. Spearman's rank correlation analysis was used to investigate the influence of elevation and surface gradient on the accumulation of the snowpack, with respect to wind influence, first for all data and subsequently for windward (north), leeward (south), east-, and west-facing slopes in isolation. This was done with reference to vegetation type using the statistical software R (version 2.12.1).

Results

Results of snow cover thickness were averaged to conform to digital map resolution. The resulting data set contains 15,826 points distributed along the studied area. Vegetation type distribution within the study area and the entire valley are approximately proportional, varying by no more than 3.2 % (Fig. 4); therefore, we consider the data set to be representative of the total study area.

Of the area covered by the GPR data set, 90.4% is vegetated. This study focuses on the hydrology of the vegetation types, and those non-vegetated land cover classes mentioned earlier were not included in these analyses and will thus not be referred to in further discussions. The obtained thickness of the snow is plotted for each vegetation type in Figure 5.

In the studied area, the degree of slope is directly proportional to the elevation, and the main valley is characterized by gentle slopes in the low-lying areas (below 400 m a.s.l.). Snow thickness normally diminishes at higher altitudes; however, this is dependent on the local topography of the terrain. Correlation analysis was used to assess the physical conditions that affect the distribution of snow for the different vegetation types (Fig. 6). In Figure 7, snow thickness is plotted against these variables.

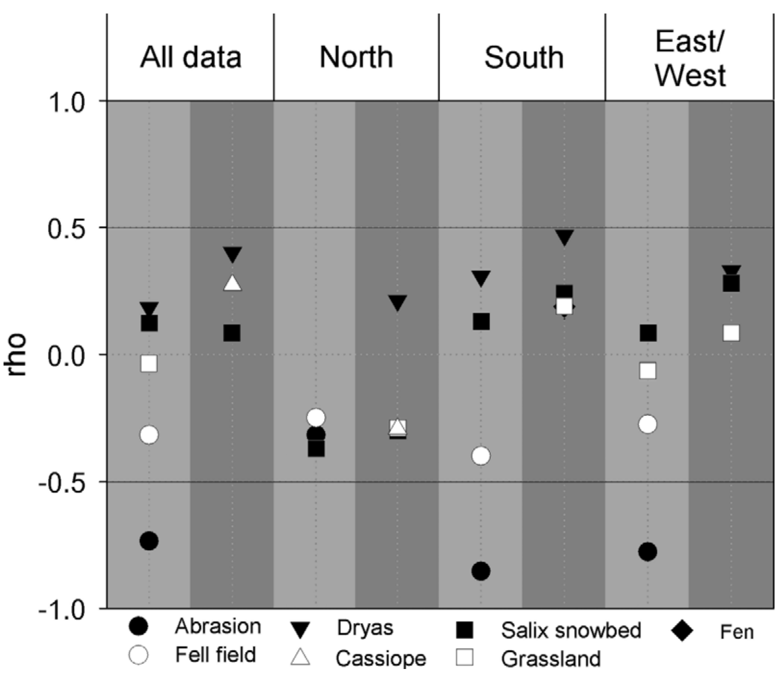

FIGURE 6. Spearman correlation coefficients for snow thickness and vegetation types. Light gray plots represent this relationship considering the influence of elevation factor. Dark gray plots represent the relationship considering the influence of the terrain slope.

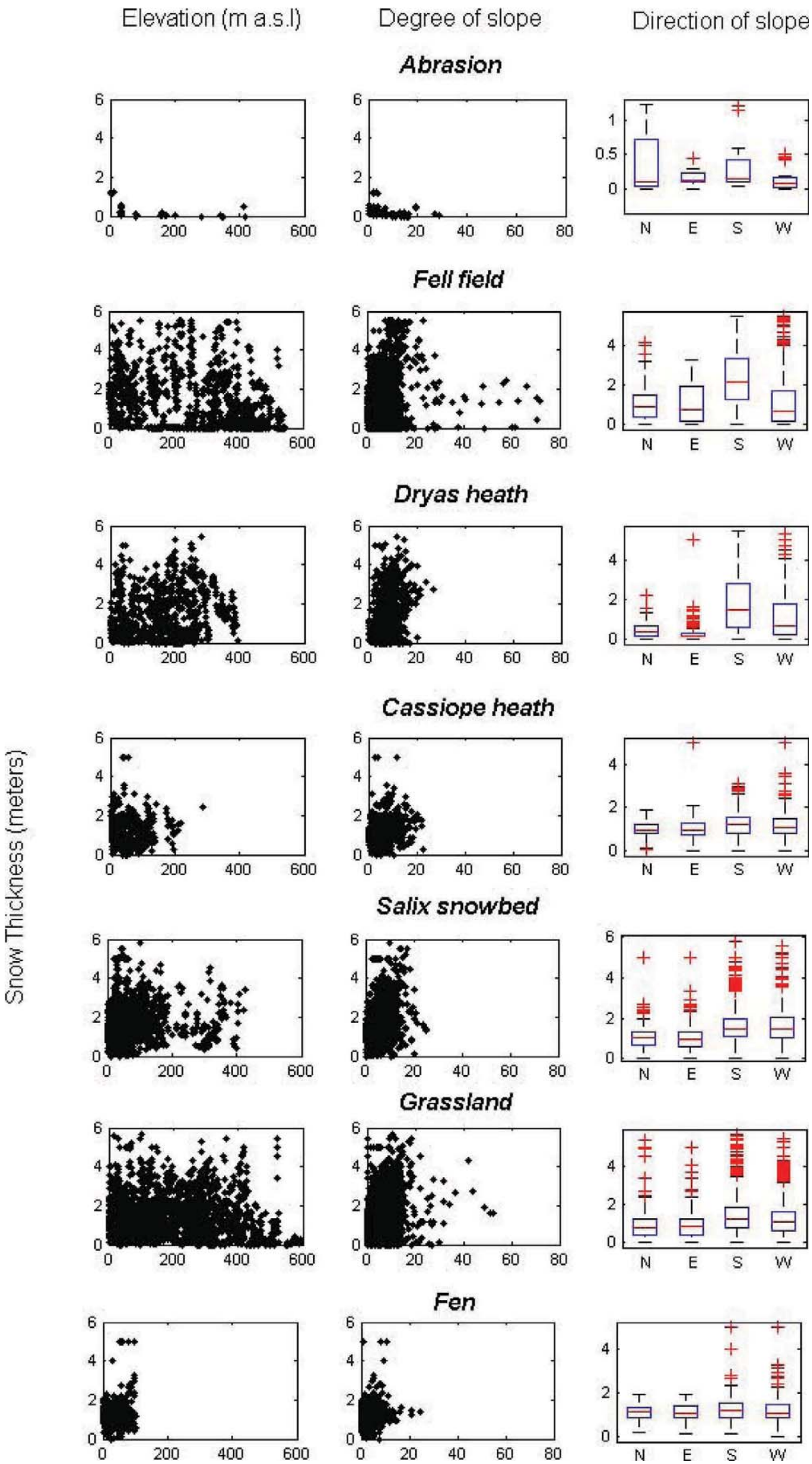

FIGURE 7. Snow thickness as a function of the physical characteristics assessed for each vegetation type.

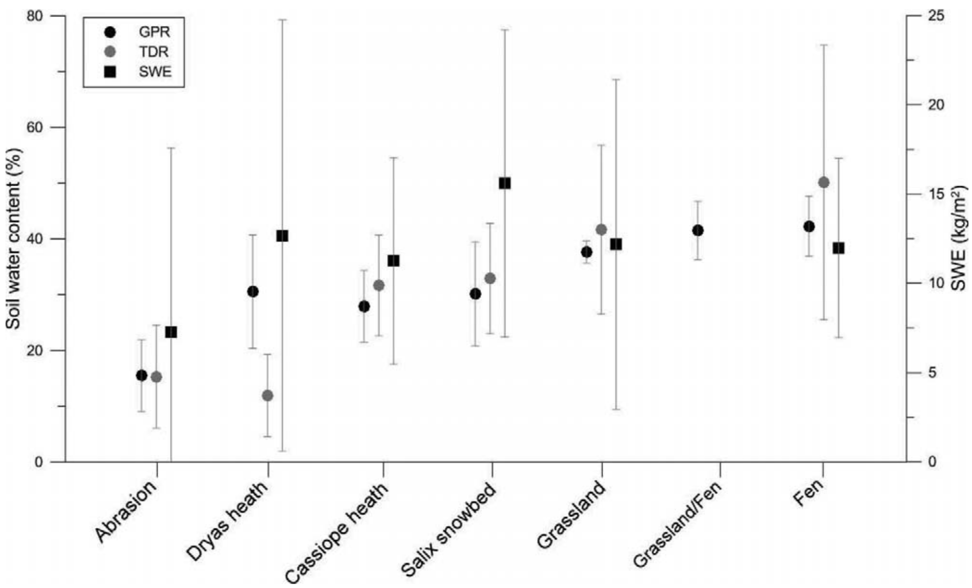

FIGURE 8. Soil saturation measured using GPR and time-domain reflectometer (TDR) at different points (left) and snow water equivalent (SWE) estimated from snowpack thickness using GPR (right), all classified by vegetation type.

SNOW DISTRIBUTION

Soil conditions make fell field and *Dryas* heath dry types of vegetation (Fig. 8). They normally have earlier snowmelt and, therefore, earlier start of the growing season (Tamstorf et al., 2007). However, both show high average snow accumulation (from Fig. 5), particularly fell field, which has higher accumulation than all other vegetation types. The occurrence of fell field and *Dryas* heath is independent of elevation, but *Dryas* heath becomes more open above 400 m a.s.l. (Fig. 7), and was classified by Bay (1998) as fell fields. On the one hand, for fell field the topographical variables have either a very low correlation coefficient (r magnitude) or are not significantly correlated to snow accumulation. On the other hand, *Dryas* heath terrain characteristics are related to the accumulation of snow, with correlation coefficients between the variables and snow stronger and significant. Leeward-facing slopes have a positive influence on snow accumulation ($r = 0.44$, $p < 0.05$), showing that for *Dryas* heath the wind has a high influence on the distribution of snow with substantially less snow accumulated in windward directions (north and east).

Abrasion plateau has extremely dry soil and is classified as a vegetation type where plants cover less than 5% (Bay, 1998). The dry condition is constrained by the amount of snow accumulated, which only averages 0.22 m (SD = 0.3 m). In this vegetation type, which is characteristic of exposed areas, snow accumulation is controlled by the wind having a clear impact on accumulation, especially for windward slopes ($r = -0.85$, $p = 0.01$).

Cassiope heath is rarely found above 150 m, hence correlations between micro-topography and snow accumulation for this vegetation type are rather weak. Consequently, the degree of slope has little influence on the homogeneity of snow accumulation with an average of 1.1 m (SD = 0.5 m). However, if we consider only areas below 50 m a.s.l., wind has no significant influence on snow redistribution. Although *Cassiope* heath is less abundant on windward slopes, snow thickness averaged between 1.0 and 1.2 m independent of slope-face orientation.

Salix snowbed is characterized by the latest snowmelt and, therefore, has a shorter growing season. Although it can also be found at high elevations, there is a predominance of *Salix* snowbed below 200 m a.s.l. There is, as expected, a clear negative correlation for windward- ($r = -0.4$, $p < 0.05$) and positive correlation for leeward-orientated slopes ($r = 0.4$, $p < 0.05$).

The ''grassland'' vegetation type is characterized by high species diversity, particularly above 200 m a.s.l. (Bay, 1998). Although significant, correlation coefficients between terrain variables and snow accumulation were weak. Below 200 m a.s.l. snow accumulation averaged 1.16 m (SD = 0.83), but for grasslands at higher elevations snow thickness averaged 1.34 m (SD = 1). Snow accumulation proved to be highly independent of the direction of slope above 200 m a.s.l.

Fens are found predominantly below 100 m a.s.l. and are defined by their high soil water content. Snow accumulation for fen (as for *Cassiope* heath) was independent of microtopography, at this scale. This vegetation type occurs where physical factors do not have a big impact on the dynamics of snow redistribution.

HYDROLOGICAL OUTCOME

When snow has melted, the active layer in the valley starts thawing and will reach a maximum depth ranging from 40 to 100 cm at the end of the summer season, depending on the soil and vegetation conditions (Christiansen et al., 2008; Gacitúa et al., 2012). Soil water content reaches its maximum immediately after snowmelt (Hasholt et al., 2008), and plant communities are highly dependent on the soil moisture, which is limited by snow thickness (Tamstorf et al., 2008). Assuming that specific petrophysical relationships and GPR data can be used to estimate water content (Slater and Comas, 2009; Gacitúa et al., 2010), a comparison between direct TDR measurements and GPR estimations in this study site are presented in Figure 8 (left *Y*-axis). TDR data were collected at different locations (ZERO line in Fig. 1), and the average of the soil moisture for each bulk point was grouped by vegetation type. The snow water equivalent (SWE) obtained from the snow thickness measurements grouped by vegetation type is also shown in Figure 8 (right *Y*-axis).

Discussion

Snow thickness measurements were conducted using geophysical methodology. The set of data contains thickness of snow from both Magnaprobe point measurements and GPR prospecting. The use of GPR relies on the accuracy of the method, which has been proven in previous research (e.g. Annan et al., 1994; Noon et al., 1998; Godio, 2008; Arcone, 2009; Previati et al., 2011). The confidence in GPR data is, however, limited by the assumption that the speed of the signal propagation in the snowpack is constant and that disturbance of the snowmobile track on the surface was of no consequence. The vertical resolution refers to the interface resolution, which is the ability to discriminate consecutive features at depth. This parameter is obtained as 75% of the wavelength (λ) (Noon et al., 1998), which is given by the rate between the signal velocity (v) and the central frequency of the antenna (f_c) ($\lambda = v/f_c$). An evaluation of the uncertainties in snow depth determination using GPR is presented by Previati et al. (2011) who, using an antenna with an approximate central frequency of 900 MHz, determined that although the theoretical vertical resolution for this antenna in snow is 0.19 m, a minimum depth of 0.3 m was required before significantly reliable results were obtained. Thus, uncertainties significantly decline for snowpack below this depth. In consequence, a proportional reliability in our data is expected. Considering that using a central frequency antenna of 500 MHz, the theoretical vertical resolution in the snow pack is 0.35 m, one would expect less uncertainty in the determination of snow thickness when the snowpack is thicker than 0.5 m. The validation plot presented in Figure 2 gives a mean difference between methods of 0.17 m (standard error of 0.036 m). The greater the depth the bigger the difference between measured and estimated values; this is probably due to either an error in reaching the real depth with the steel rod or an error in the determination of the interface of snow/soil in the radar data due to noise and scatter of the signal.

From our total set of snow thickness data, 94% correspond to GPR measurements and from this only 17% are measurements of thickness below 0.5 m. Therefore, we could expect high accuracy for most of the data. However, for those vegetation types that are considered dry and where snow accumulation is in general low such as Abrasion, *Dryas*, and fell field, values below 0.5 m of snow thickness measured using GPR are 58%, 30%, and 18%, respectively, of the data available for each type. This suggests a proportionally higher uncertainty in estimations that are the outcome of GPR data for these vegetation types.

Despite the accuracy of GPR data it can be argued there is a wide advantage on the use of this method because it provides continuous data of the subsurface with minimal disturbance. In contrast, when collecting point measurements (Magnaprobe), data is limited in depth and spatial resolution. The same condition applies to soil moisture estimation from GPR and direct measurements using TDR. On one hand, given the several assumptions involved for GPR, estimations of water content can yield a high uncertainty (Gacitúa et al., 2012). On the other hand, TDR measurements are restricted to point measurements, in this case values that are averages of measurements taken at 35 points at increments of either 5 or 10 cm (40 cm maximum depth), which gives limited resolution of the total study area.

Studies in Low Arctic regions show that snow depth correlates closely with shrub canopy height and stem diameter in Arctic vegetation, as copse and bush tend to capture drifting snow (McFadden et al., 2001). However, at High Arctic latitudes such as in the Zackenberg valley, the average height for all plant communities (except for fen) typically does not exceed 7 cm (Bay, 1998). Consequently, with an average snow depth of 0.80 m during winter, vegetation does not influence snow accumulation in this area. Instead, plant communities are themselves affected by the amount of snow accumulated, which influences the length of the growing season. A large interannual variability in the amount of snow accumulated in the Zackenberg valley by the end of the winter has been found through prior studies and monitoring—e.g. $0.17 - 1.33$ m during 1997–2010 (Hinkler et al., 2008; Sigsgaard et al., 2011). This results in large variation in the length of the snow-free growing season, which is considerably shorter at higher latitudes. This is particularly important for the distribution of *Cassiope* heath and *Salix* snowbed. According to the conceptual gradient in the distribution of plant communities with altitude as depicted by Elberling (2008), there is a clear transition (down slopes) from *Cassiope* heath to *Salix* snowbed. This characteristic influences the amount of water available for the soil water transfer from *Cassiope* heath to *Salix* snowbed during spring and summer. Measurements show that there is an average of 0.42 m more snow accumulated in *Salix* snowbed than in *Cassiope* heath (Fig. 5) which leads to increased water availability during the growing season and a reduction in the length of the snow-free growing season.

Snow accumulation is highly influenced by the slope and elevation of *Dryas* heath and *Salix* snowbed, and therefore also by the direction of the wind due to the fact that snow accumulation is significantly lower on windward than on leeward slopes. Snow accumulation in *Cassiope* heath, grassland, and fen showed, in general, weak or absent dependence on the physical conditions for which they were evaluated.

Soil water content drops as a consequence of evaporation and drainage during the spring and summer. Snow ablation processes have not been considered in the estimation of the final water outcome of the snowpack; however, Figure 8 shows that using GPR for estimating snow water equivalence of the winter snowpack and soil moisture during the summer produces proportional results to those obtained using TDR during spring.

Summary

The methodology used in this study consisted of collection of site measurements of snow thickness (and density) to assess the possible relationships among the factors that can influence snow accumulation, such as terrain gradient, elevation, and wind direction. This was done in relation to the vegetation types, assuming the strong dependence that they have on the hydrological input, which they have from the snowpack thickness and its consequent melt.

Results from correlation analyses showed that although wind direction had an influence on snow redistribution in the valley, this was nonsignificant for some vegetation types, especially those that dominate lowland areas (e.g. grassland and fen), whereas neither elevation nor slope influenced snowpack thickness significantly. Nevertheless, the thickness of accumulated snow on vegetation types that dominate the uplands (e.g. abrasion and fell field) showed a significant negative correlation with elevation. It should, however, be noted that fell fields were uniformly distributed within the valley, and snow thickness was relatively homogeneous in relation to elevation. The correlation coefficient was in most cases significant, due to the large size of the data set. However, significance was generally weak for most of the correlations assessed.

As for most of the High Arctic region, the future winter climate of Northeast Greenland is predicted to become considerably warmer with increased precipitation, but also more variable, both temporally and spatially (Anisimov et al., 2007). The results presented here can be useful in future studies that compare variation in local snow accumulation and vegetation distribution. They could also be used in the assessment of snow-vegetation interaction and for quantifying the potential impact of increasingly variable precipitation on vegetation and the associated hydrological processes that snow produces in the soil.

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