

Anisotropy of seasonal snow measured by polarimetric phase differences in radar time series

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Abstract. The snow microstructure, i.e. the spatial distribution of ice and pores, generally shows an anisotropy which is driven by gravity and temperature gradients and commonly determined from stereology or computer tomography. This structural anisotropy induces anisotropic mechanical, thermal, and dielectric properties. We present a method based on radio wave birefringence to determine the depth-averaged, dielectric anisotropy of seasonal snow with radar instruments from space, air or ground. When snow depth and density are known, the birefringence allows determination of the dielectric anisotropy by measuring the copolar phase difference (CPD) between linearly polarized microwaves propagating obliquely through the snowpack. The dielectric and structural anisotropy are linked by Maxwell–Garnett-type mixing formulas. The anisotropy evolution of a natural snowpack in Northern Finland was observed over four winters (2009–2013) using the ground-based radar instrument "SnowScat". The radar measurements indicates horizontal structures for fresh snow and vertical structures in old snow which is confirmed by computer tomographic in-situ measurements. The ground based CPD data were compared with space-borne measurements from the satellite TerraSAR-X which showed the same temporal evolution. The presented dataset provides a valuable basis for the development of new snow models which include the anisotropy of the snow microstructure.

1 Introduction

After deposition on the ground, snow crystals form a porous, sintered material which continuously undergoes metamor-

phism to adapt to the thermodynamic forcing imposed by the atmosphere and the soil. The porous microstructure, defined by the 3D distribution of the ice matrix and the pores space, determines the thermal, mechanical and dielectric properties of the snowpack. Hence, a spatially anisotropic distribution of the microstructure leads to a macroscopic anisotropy of snow properties.

Characterization of the microstructure is difficult and requires work intensive sampling, sample preparation, and data processing but enables a unique insight into the structure at micrometer scales. Macroscopic (point) methods commonly applied in the field can be used to determine snow properties averaged over sample volumes of several centimeters. Methods based on remote sensing complement these point methods in providing large spatial coverage of repetitive measurements with a sampling resolution between meters and kilometers also for inaccessible locations. For snow, radar remote sensing methods facilitates measurements of snow properties averaged over the microwave penetration depth. This makes it possible to estimate the depth-averaged dielectric anisotropy of seasonal snow with radar instruments.

1.1 Observations of the structural anisotropy

Anisotropic structures have been identified in photographs of thin section cuts of seasonal snow with preferentially horizontal structures for fresh snow and vertical structures in old snow (Kojima, 1960; Davis and Dozier, 1989; Mätzler, 1987, Fig. 2.15¹). Vertical structures have been reported also in thin

¹Note that the captions of Figs. 2.14 and 2.15 in Mätzler (1987) have been inadvertently swapped.

sections cuts of polar firn (e.g. Alley, 1987). The formation of anisotropic, vertical snow structures has been observed by thin section photography, when snow metamorphism was driven by a vertical water vapor flux under temperature gradients (e.g. Pfeffer and Mrugala, 2002).

The anisotropy of snow can be statistically determined from the snow microstructure by spatial correlation functions (e.g. Vallese and Kong, 1981; Mätzler, 1997) computed from stereological (Alley, 1987; Mätzler, 2002, and others) or computer tomography data (e.g. Löwe et al., 2011, 2013). Today, computer tomography is often considered as the “state of the art” for destruction-free observations of the snow microstructure within volumes of a few cm^3 and with a spatial resolution on the micrometer-scale. The non-destructive CT measurements allow imaging of the same sample multiple times to observe the temporal evolution of the microstructure for samples kept under laboratory conditions (Schnee-
beli and Sokratov, 2004). The formation of vertical structures from initially horizontal structures has been observed in laboratory samples if vertical temperature gradients are applied (Schnee-
beli and Sokratov, 2004; Riche et al., 2013; Calonne et al., 2014). Vertical as well as horizontal structures have been found in artificial snow from the cold-laboratory but also in natural seasonal snow (Calonne et al., 2012). In polar firn, vertical structures are commonly found in firn cores at different depths (e.g. Hörhold et al., 2009; Fujita et al., 2009; Lomonaco et al., 2011).

In contrast to vertical structures which are known to be caused by vertical temperature gradients, horizontal structures have a different origin. A horizontal orientation is initially created by the deposition of anisotropic, atmospheric growth forms (plates, needles, dendrites) which predominantly align horizontally in the gravity field (Matrosov et al., 2005; Garrett et al., 2012). Snow settling can further contribute to horizontal anisotropies in the intermediate stage of isothermal metamorphism (Löwe et al., 2011), where only gravity breaks the symmetry between vertical and horizontal directions.

1.2 Field observations of the dielectric anisotropy

On macroscopic scales, the anisotropy of snow can be characterized by measuring the anisotropy of the thermal conductivity (e.g. Izumi and Huzioka, 1975) or of the dielectric permittivity. The anisotropy of the thermal conductivity is much stronger compared to the anisotropy of the dielectric permittivity, because the contrast in thermal conductivity k between ice and air $k_{\text{ice}}/k_{\text{air}}$ is ≈ 100 compared to the contrast in permittivity between ice and air which is $\varepsilon_{\text{ice}}/\varepsilon_{\text{air}} \approx 3$ (Löwe et al., 2013). Still, dielectric measurements have been discussed already in 1965 with respect to the shape and orientation of ice crystals (e.g. Evans, 1965).

The dielectric anisotropy, $\Delta\varepsilon$, in the following defined as the difference between the horizontal and vertical permittivities, $\Delta\varepsilon = \varepsilon_x - \varepsilon_z$, can be measured precisely using dif-

ferent polarizations of the electromagnetic field. The dielectric anisotropy in ice, caused by the c -axis orientation of the crystal fabric, has been measured with open microwave resonators (Matsuoka et al., 1996, 1997) of the design of Jones (1976). With the same method, the dielectric anisotropy in snow, caused by a structural anisotropy of the ice matrix, has been measured and higher permittivities have been found in the vertical direction compared to the horizontal direction in multi-year firn on both, the Greenland ice sheet (Fujita et al., 2014) and on the Antarctic ice sheet (Fujita et al., 2009, 2016); Fujita et al. (2009) did the analysis in conjunction with a computer tomographic analysis. Using a method of microwave propagation, Lytle and Jezek (1994) also detected larger a vertical than horizontal permittivity in multi-year firn on the Greenland ice sheet combined with a photographic analysis. Sugiyama et al. (2010) found similar results in Antarctica: the measured horizontal permittivity in the upper 1m snow layer were often smaller than expected from empirical relations between permittivity and density of isotropic snow.

1.3 Radio and microwave remote sensing observations of the dielectric anisotropy

Polarimetric radar remote sensing methods can provide information of the dielectric anisotropy of snow and ice from large distances. Areas of many thousands of square-kilometers can be observed with air- and space-borne sensors, even repeatedly by satellites during every repeat-pass of an orbit. The imaging resolution and penetration depth of radio- or microwaves implies that usually area-, depth- or volume-averaged snow properties are measured. Remote sensing methods provide therefore a complementary tool to the detailed ground measurements such as computer tomography or in-situ measurements.

For example, measurements of the radio wave birefringence have been used to explore internal structures of ice sheets and glaciers with polarized radio- and microwaves (e.g., Hargreaves, 1977, 1978; Fujita et al., 2006; Matsuoka et al., 2009; Parrella et al., 2015). Also with passive microwave sensors, strong polarimetric signatures have been found over the Greenland ice sheet: in Li et al. (2008), the observed passive microwave signatures could not be explained by surface features, but were discussed with respect to microstructural variations of the anisotropy of snow, as predicted by Tsang (1991).

For seasonal snow, the observation of both the structural anisotropy and the dielectric anisotropy in polar firn indicates that seasonal snow should also be a dielectrically anisotropic medium. Still, publications related to polarimetric propagation effects in deposited seasonal snow are rare, despite the fact that a differential propagation speed in falling snow was already noticed in 1976 for weather radars (Hendry et al., 1976). Today, polarimetric upwards looking radars are used to characterize the orientation and anisotropy of falling snow

particles or rain (e.g. Matrosov et al., 2005; Garrett et al., 2012; Xie et al., 2012; Hogan et al., 2012; Noel and Chepfer, 2010; Tyynelä and Chandrasekar, 2014).

The dielectric anisotropy of seasonal snow can be determined by analyzing the propagation differences of differently polarized microwaves propagating obliquely through the snowpack. In contrast to the observations for firm on ice sheets where $\varepsilon_z > \varepsilon_x$, a larger horizontal permittivity ε_x was found in freshly deposited snow using ground-based radar measurements: Chang et al. (1996) measured the propagation difference between vertically and horizontally polarized microwaves by measuring their phase difference, the so called Copolar Phase Difference (CPD), and found that the CPD increased after snow fall. The increase of the CPD was explained by a "horizontal alignment of new snow crystals" (Chang et al., 1996). In polarimetric measurements acquired by the radar satellite TerraSAR-X (Stangl et al., 2006; Werninghaus and Buckreuss, 2010), both, positive and negative CPDs were observed at different times by Leinss et al. (2014b): a positive correlation between the CPD and the depth of fresh snow was found, indicating horizontal structures in fresh snow where $\varepsilon_x > \varepsilon_z$. They also observed the opposite effect, where a strong temperature gradient in the snowpack forced the CPD towards negative values, indicating vertical structures where ($\varepsilon_x < \varepsilon_z$).

1.4 Paper structure

In this paper, we present and apply an electromagnetic model to determine the depth-averaged structural anisotropy of a dry snowpack via the dielectric anisotropy derived from the CPD measured with polarimetric radar systems. The paper is structured as follows:

- Sect. 2 revisits Maxwell-Garnett mixing formulas and also links them to the microstructural characterization of snow in terms of spatial correlation functions to calculate the anisotropic dielectric permittivity based on the structural anisotropy.
- Sect. 3 derives the CPD measured by polarimetric radar systems for an anisotropic dielectric medium of known thickness and negligible scattering and absorption losses.
- Sect. 4 describes the experiment and the field data collected within four winter seasons from 2009 to 2013. The CPD measurements are discussed in view of marker events like snow fall, snow metamorphism and melting.
- Sect. 5 analyzes the CPD time series to determine the evolution of the depth-averaged anisotropy of the snowpack. For four selected dates, the derived anisotropy is compared with measurements obtained by computer-tomography. Furthermore, we compare the measured CPD time series with space-borne radar observations

from TerraSAR-X and discuss whether the CPD can be used to determine the depth of fresh snow.

- Sect. 6 concludes the paper.

2 Dielectric permittivity linked to structural anisotropy

The ice matrix of snow can have two different anisotropies which both can influence the anisotropy of the dielectric permittivity. The structural anisotropy, discussed in this section, is given by an anisotropic spatial distribution of the ice matrix i.e. the shape of ice crystals. The crystal fabric anisotropy, discussed in Appendix A, is given by the crystal axis orientation of ice crystals but its effect on the permittivity of seasonal snow is small compared to the effect of the structural anisotropy.

In this section an adapted version of the Maxwell-Garnett mixing formulas is presented to provide a relation between the structural anisotropy of snow, snow density and the effective dielectric permittivity ε_{eff} .

In the following we define the coordinate axes such that z is the vertical (parallel to gravity) and the $x - y$ plane is parallel to the horizontal earth surface. We restrict our model to flat terrain and do not consider the possibility that the symmetry axis of the snow structure is not parallel to gravity, which can be the case in steep terrain.

2.1 Definition of structural anisotropy

We define the structural anisotropy, A , as the normalized difference between the characteristic horizontal dimension, a_x , and the characteristic vertical dimension, a_z , of the "grains" in the ice matrix:

$$A = \frac{a_x - a_z}{\frac{1}{2}(a_x + a_z)} \quad (1)$$

Different choices for the length scales a_x and a_z are possible (Löwe et al., 2011). Recent work for microwave modeling has mainly used the (exponential) correlation lengths, $a_x = p_{\text{ex},x}$ and $a_z = p_{\text{ex},z}$, as defined in Mätzler (2002). The exponential correlation lengths are conveniently derived by an exponential fit to spatial correlation functions (Löwe et al., 2013).

The advantage of the normalized difference, Eq. (1), is that A only changes sign but not magnitude if the orientation of the longest length changes its orientation from vertical to horizontal (while keeping a fixed ratio between longest and shortest length). For the common definition A' , where the anisotropy is defined by the length ratio a_z/a_x , the magnitude of the difference to the isotropic case ($A'_{\text{iso}} = 1$) depends on the orientation of the longest length: the difference becomes clear when comparing e.g. ($a_x = 2$, $a_z = 1$), yielding $A = 0.66$ and $A' = 0.5 = A'_{\text{iso}} - 0.5$ with ($a_x = 1$, $a_z = 2$) which gives $A = -0.66$ and $A' = 2 = A'_{\text{iso}} + 1.0$. The

definition A' is commonly used to simplify electromagnetic modeling. The anisotropy A' can be converted to Eq. (1) by

$$A' = \frac{a_z}{a_x} = \frac{2 - A}{2 + A} \quad \text{or} \quad A = \frac{1 - A'}{\frac{1}{2}[1 + A']}. \quad (2)$$

We note that the anisotropy A' differs from the definition in terms of the “degree of anisotropy” (DA) which is used in (Hildebrand et al., 1999; Schneebeli and Sokratov, 2004). For the DA, the absolute orientation in space is lost since the definition is based on the ratio of the largest and smallest eigenvalues of the mean intersection length (MIL) tensor. The anisotropy A' (defined as ϵ in Torquato and Lado, 1991 or $\mathcal{A}(l_c)$ in Calonne et al., 2014) can be further related to the anisotropy-parameter Q used in Calonne et al. (2014) by the definition in Löwe et al. (2013, Eq. 4).

2.2 Relative permittivity as a function of anisotropic inclusions: Maxwell-Garnett formulas

The CPD measured by polarimetric radar systems depends on the difference of the dielectric permittivity ϵ_{eff} measured in the x and z direction. The aim of this subsection is to establish a link between the effective permittivity $\epsilon_{\text{eff},i}$ for $i \in \{x, y, z\}$ and the structural anisotropy A . The following model is based on an empirical extension of the classical Maxwell–Garnett mixing formulas for aligned mixtures of ice inclusions in a host medium of air (e.g. Polder and van Santen, 1946; Sihvola, 2000).

To motivate the necessity of the empirical extension we briefly revisit the application of Maxwell–Garnett mixing formulas in the isotropic case. For isotropic snow ($A = 0$) the modeled permittivity $\epsilon_{\text{eff},i}$ should agree with measurements of ϵ of isotropic snow. However, the relative permittivity, $\epsilon_{\text{eff, MG}}$, calculated with the Maxwell–Garnett formula underestimates the measured permittivity (Mätzler, 1996) which shows slightly higher values. The reason is that $\epsilon_{\text{eff, MG}}$ is equivalent with the lower Hashin–Shtrikman bound (Sihvola, 2002; Hashin and Shtrikman, 1962). The upper Hashin–Shtrikman bound is equivalent with the “inverse” Maxwell–Garnett formula, $\epsilon_{\text{eff, MG, inv}}$, which models air inclusions in a host medium of ice (Sihvola, 2002). Therefore it is preferable to combine both bounds in a reasonable way to determine ϵ_{eff} . We found that the following weighted average,

$$\epsilon_{\text{eff}} = (\epsilon_{\text{eff, MG}} + \epsilon_{\text{eff, MG, inv}} \cdot f_{\text{vol}} \epsilon_{\text{ice}}) / (1 + f_{\text{vol}} \epsilon_{\text{ice}}), \quad (3)$$

agrees best with values from literature (for details see Appendix B). The ice volume fraction f_{vol} relates the density of snow ρ (g/cm^3) to the volumetric mass density of air and ice by

$$\rho = f_{\text{vol}} \cdot \rho_{\text{ice}} + (1 - f_{\text{vol}}) \cdot \rho_{\text{air}} \approx f_{\text{vol}} \cdot \rho_{\text{ice}}. \quad (4)$$

In the microwave regime between 10–20 GHz, the relative permittivity of pure polycrystalline ice is given by $\epsilon_{\text{ice}} =$

3.17 ± 0.02 , and shows a weak temperature dependence (Mätzler and Wegmüller, 1987; Fujita et al., 1993; Matsuoka et al., 1996; Warren and Brandt, 2008; Bohleber et al., 2012). As the uncertainty for snow density measurements of some percent is larger than the temperature dependence of the permittivity, a fixed permittivity $\epsilon_{\text{ice}} = 3.17$ is used in this paper, corresponding to a temperature of about -10°C .

According to the Maxwell–Garnett theory for isotropic mixtures, $\epsilon_{\text{eff, MG}}$ is given by

$$\epsilon_{\text{eff, MG}} = \epsilon_{\text{air}} + 3f_{\text{vol}} \epsilon_{\text{air}} \frac{\epsilon_{\text{ice}} - \epsilon_{\text{air}}}{\epsilon_{\text{ice}} + 2\epsilon_{\text{air}} - f_{\text{vol}}(\epsilon_{\text{ice}} - \epsilon_{\text{air}})} \quad (5)$$

with the relative permittivity of air, $\epsilon_{\text{air}} = 1$ (e.g. Sihvola, 2000). The “inverse” Maxwell–Garnett result, $\epsilon_{\text{eff, MG, inv}}$, follows by swapping ϵ_{air} and ϵ_{ice} in Eq. (5) and replacing f_{vol} by $1 - f_{\text{vol}}$ (Sihvola, 2002). Note that the Maxwell–Garnett theory is a mean-field theory and additionally requires the inclusions to be much smaller than the wavelength of the microwaves in the medium ($a_x, a_y, a_z \ll \lambda / \sqrt{\epsilon_{\text{eff}}}$), so that scattering in the snow volume can be neglected.

For non-spherical inclusions, Eq. (5) has to be adapted by introducing depolarization factors, N_i , for aligned ellipsoidal inclusions (e.g. Cohn, 1900; Polder and van Santen, 1946, or Sihvola, 2000). As settling and temperature gradient metamorphism act in the z direction, we model the elliptical inclusions as oblate or prolate spheroids which have their symmetry axis parallel to z . According to Sihvola (2000) the permittivity of anisotropic mixtures is given for each spatial dimension $i \in x, y, z$ by

$$\epsilon_{\text{eff, MG}, i} = \epsilon_{\text{air}} + f_{\text{vol}} \epsilon_{\text{air}} \frac{\epsilon_{\text{ice}} - \epsilon_{\text{air}}}{\epsilon_{\text{air}} + (1 - f_{\text{vol}}) N_i (\epsilon_{\text{ice}} - \epsilon_{\text{air}})} \quad (6a)$$

The “inverse” Maxwell–Garnett form of Eq. (6a) reads

$$\epsilon_{\text{eff, MG, inv}, i} = \epsilon_{\text{ice}} + (1 - f_{\text{vol}}) \epsilon_{\text{ice}} \frac{\epsilon_{\text{air}} - \epsilon_{\text{ice}}}{\epsilon_{\text{ice}} + f_{\text{vol}} N_i (\epsilon_{\text{air}} - \epsilon_{\text{ice}})}. \quad (6b)$$

Both equations are used in Eq. (3) to calculate the effective anisotropic relative permittivities of snow, $\epsilon_{\text{eff}, x}$ and $\epsilon_{\text{eff}, z}$. Results for the permittivity and the deviation from the isotropic case are shown in Fig. 1.

The depolarization factors N_i are assumed to be equivalent for both Eqs. (6a) and (6b) as both equations describe the polarizability of elliptical particles. The depolarization factors N_i are given according to (Sihvola, 2000) for ellipsoidal inclusions with the dimensions a_x, a_y, a_z by the elliptic integral of second kind

$$N_i = \frac{a_x a_y a_z}{2} \int_0^\infty \frac{ds}{(s + a_i^2) \sqrt{(s + a_x^2)(s + a_y^2)(s + a_z^2)}}, \quad (7)$$

where the integration variable s (units: distance square) describes an ellipsoidal surface larger than the surface of the elliptic inclusion on which $s = 0$ (Landau and Lifshitz, 1960,

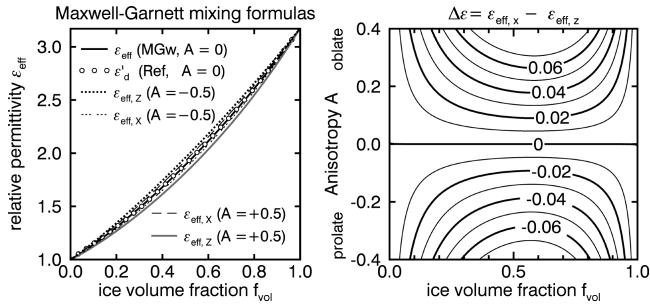


Figure 1. Left: Relative permittivity ϵ_{eff} of snow with isotropic ($A = 0$, solid), vertically oriented ($A = -0.5$, dots) and horizontally oriented ($A = +0.5$, dashed) inclusions calculated by the weighted Maxwell–Garnett formula (MGw), Eq. (3). The open circles indicate the empirical function given by Eq. (46) in Wiesmann and Mätzler (1999). Right: the dielectric anisotropy, $\Delta\epsilon = \epsilon_{\text{eff},x} - \epsilon_{\text{eff},z}$, as a function of ice volume fraction f_{vol} and anisotropy A according to Eq. (3).

§4, p20-30). The dimensions $a_x = a_y$ define the (horizontal) diameter of the spheroids and a_z is their vertical length. Note that Sihvola (2000) used the ellipsoids’ semi-axis. However, the depolarization factors do not depend on the absolute size of inclusions and are invariant under rescaling $a_i \rightarrow \lambda a_i$ for arbitrary λ . Consequently, it is possible to parametrize the depolarization factors directly by the anisotropy A' , which can easily be verified by substituting s in Eq. (7) with the dimensionless quantity $u = s/a_x^2$. N_i is then given by

$$N_i = \frac{A'}{2} \int_0^{\infty} \frac{du}{(u + \delta_{A'}(i, z)) \sqrt{(u+1)^2 \cdot (u + A'^2)}} \quad (8)$$

with $\delta_{A'}(i, z) = 1$ for $i \in x, y$ and $\delta_{A'}(i, z) = A'^2$ for $i = z$. Closed form expressions for the elliptic integrals can be found e.g. in Landau and Lifshitz (1960); Sihvola (2000). The depolarization factors satisfy $N_x + N_y + N_z = 1$ for any ellipsoid (Polder and van Santen, 1946). For spherical inclusions all three depolarization factors are $N_i = 1/3$ and Eq. (6a) is equivalent with Eq. (5).

2.3 Series expansion of permittivity from spatial correlation functions: equivalence with Maxwell-Garnett formulas

Although ice grains show a much more complex structure than simple ellipsoids, the model of ellipsoids is realistic enough for the transverse isotropic symmetry of the dielectric tensor $\bar{\epsilon}$. This becomes more obvious from the exact series expansion of the dielectric tensor for arbitrary anisotropic microstructures, which can be expressed in terms of spatial correlation functions (Rechtsman and Torquato, 2008). In Appendix C, we show that under the less restrictive assumption of a transverse isotropic two-point correlation function, the truncation of the exact expression using n -point correlation functions (Rechtsman and Torquato, 2008, Eq. 16) at

second order ($n = 2$) exactly leads to the Maxwell–Garnett result (Eq. 6a) in which the depolarization factors N_i are expressed in terms of the anisotropy parameter Q as given in (Löwe et al., 2013) via $N_i = Q$ for $i = x, y$ and $N_z = 1 - 2Q$. This implies that the present dielectric model and the thermal conductivity model from (Löwe et al., 2013) are based on exactly the same microstructural parameters. In view of recent attempts to unify microstructural descriptions of snow for microwave modeling (Löwe and Picard, 2015), we also note that the Maxwell–Garnett formula (Eq. 6a) can be likewise obtained as the low-frequency limit of the quasi-crystalline approximation for aligned spheroids (Ao and Kong, 2002).

3 Dielectric anisotropy measured by polarimetric radar systems

The depth-averaged anisotropy of a snowpack of known depth and density can be measured with an obliquely looking polarimetric radar system when the phase difference between two perpendicular polarized microwaves pulses reflected at the bottom of the snowpack is analyzed with respect to the microwave-birefringence of snow.

3.1 Experimental considerations

For measuring quantitatively the dielectric anisotropy of the snowpack, the angle between the field vector of the electromagnetic field and the principal axes of the dielectric tensor $\bar{\epsilon}$ must be known. Two orthogonally polarized microwaves should be chosen such that the polarizations are delayed by different components of the dielectric tensor $\bar{\epsilon}$. The anisotropy of seasonal snow has the symmetry axis in the vertical, therefore one polarization must be at least partially aligned with the vertical while the other polarization must be oriented horizontally. Side-looking polarimetric radar systems like real or synthetic aperture radar systems using a vertical (VV) and horizontal (HH) polarization as defined in Fig. 2 fulfill this requirement, whereas the anisotropy cannot be measured by nadir-looking radar systems (e.g. ground penetrating radars) as long as there is no anisotropy in the horizontal (xy-) plane.

A further requirement is that the depth where most of the microwave energy is backscattered is known. For dry snow and frequencies of a few GHz where the volume scattering contribution in shallow seasonal snow is negligible (e.g. Hallikainen et al., 1987; West et al., 1993; Tsang et al., 2007, or Leinss et al., 2015, Fig. 5), this requirement is easy to fulfill and the scattering center corresponds to the soil below the snowpack. However, for deep firm on ice sheets or glaciers it can be difficult to obtain a good estimate on the penetration depth. The following method is not suitable for wet snow, as the dielectric properties, especially absorption and the penetration depth, strongly depend on the water content.

3.2 CPD definition and signal-processing basis

The Copolar Phase Difference (CPD) can be used for measuring for phase difference resulting from different propagation speeds of two orthogonally polarized microwaves. The CPD is defined as the phase difference between the complex-valued backscatter coefficients, S_{VV} and S_{HH} . The scattering coefficient S_{VV} for the V polarization (V transmit, V receive) is determined by the coherent superposition of all scattered fields from the ensemble of scatterers contained in the corresponding range resolution cell of the radar (the pixel). The random distribution of scatterers in the resolution cell defines the unknown but deterministic scattering-phase ϕ_{VV} of the coefficient S_{VV} . The scattering coefficient S_{HH} is equivalently defined for perpendicular transmit (H) and receive polarization (H). If the scatterers do not have a polarization dependent scattering phase and if no birefringent medium exists between the radar and the scatterers, the (generally random) phase values ϕ_{VV} and ϕ_{HH} are spatially correlated and show a zero-phase difference. The (complex-valued) spatial correlation function is defined by the copolar coherence

$$\gamma_{VV,HH} \cdot e^{i\phi_{CPD}} = \frac{\langle S_{VV} \cdot S_{HH}^* \rangle}{\sqrt{\langle |S_{VV}|^2 \rangle \cdot \langle |S_{HH}|^2 \rangle}}. \quad (9)$$

The magnitude of the copolar coherence defines the correlation coefficient $\gamma_{VV,HH}$ ranging from 0 to 1. The notation $\langle \cdot \rangle$ indicates a spatial average over about 10 to several thousands of pixels containing the backscatter coefficients of each polarization and the asterisk * denotes complex conjugation. The CPD is equivalent to the phase of the copolar coherence,

$$\phi_{CPD} \approx \langle \phi_{VV} - \phi_{HH} \rangle. \quad (10)$$

For monostatic radar systems, the same coordinate system (H, k , V, Fig. 2) is used for transmission and reception of the microwave signal, which is called “Back-Scatter Alignment” convention, BSA, (cf. Lüneburg and Boerner, 2004 or Lee et al., 1999, Sect. 3.1.3) and which reverses the wave vector k in the receiving coordinate system. The reversal of k in the BSA causes a sign-change of the CPD, hence the physically expected phase difference caused by birefringent media ϕ'_{CPD} is related to the phase difference measured in the BSA by

$$\phi_{CPD} = (-1) \cdot \phi'_{CPD}. \quad (11)$$

The magnitude of the coherence is reduced when two corresponding resolution cells (of same range but different polarization) contain scatterers which do not show a correlation between orthogonal polarizations. This is the case for objects showing strong multiple scattering (e.g. rough surfaces and strongly scattering volumes). The coherence is also reduced when the corresponding range-resolution cells, represented by S_{VV} and S_{HH} , are not perfectly overlapping and do therefore not contain exactly the same ensemble of scatterers. This occurs e.g. for large propagation delays ΔR between the two polarizations. For partially overlapping resolution cells of size δr , the coherence is reduced proportional

to $1 - \Delta R/\delta r$. The coherence is totally lost when ΔR exceeds the range resolution δr of the radar system. For partially overlapping resolution cells, only scatterers which are both contained in the resolution cell of different polarizations contribute constructively to the coherence; other scatterers lead to decorrelation. The contribution of correlated scatterers to the CPD can therefore be described by two polarized waves which have a common wave front before propagating through a birefringent medium and which are scattered at exactly the same point P on the ground. This scattering geometry is the basis of Fig. 2.

The dielectric anisotropy can precisely be measured with the CPD, because the CPD, defined as the phase of a signal, can be determined with a precision of a few degrees (i.e. fraction of one wavelength) relative to the total phase delay of many wavelengths which is accumulated during propagation through the snowpack (Gunteriusen et al., 2001, Eq. 5), and (Leinss et al., 2015, Eq. 14). For example, for 1 m deep snow of density $\rho = 0.25 \text{ g cm}^{-3}$ a dielectric anisotropy $\Delta\epsilon = \epsilon_x - \epsilon_z = 10^{-4}$ causes a CPD of 1° relative to the total phase delay of 5700° measured at a radar frequency of 10 GHz and a radar incidence angle of 40° . The dielectric anisotropy of transparent media (e.g. a dry snowpack) can therefore be measured much more accurately with the CPD compared to the time-delay between two perpendicularly polarized microwave pulses.

3.3 CPD of birefringent, non-scattering media

In order to derive the CPD, the wave propagation through snow is formulated in analogy to transversely isotropic media as done in anisotropic optics (Saleh and Teich, 1991). Considering snow as transversely isotropic is reasonable since gravity and the direction of the water vapor flux in snow break isotropy in the vertical direction, therefore the optical axis is given by the z axis.

According to anisotropic optics, we define the refractive index in the z direction as the extraordinary refractive index n_e . For transversely isotropic media, the extraordinary refractive index, n_e , differs from the ordinary refractive indices, n_o , which is defined in the (x, y) plane (Fig. 2). The refractive indices are related to the relative permittivity defined in Eq. (3) together with Eqs. (6a) and (6b) by

$$n_o^2 = \epsilon_{\text{eff},x} = \epsilon_{\text{eff},y} \quad (12a)$$

$$n_e^2 = \epsilon_{\text{eff},z}. \quad (12b)$$

The anisotropy of snow can only be determined with polarimetric radar systems when microwaves are transmitted with a large enough incidence angle θ_0 with respect to the optical axis. The polarizations of a side-looking radar system are defined orthogonal to the propagation vector k of the incident beam such that the horizontal polarization (H) is oriented parallel to the observed surface (cf. Fig. 2). Hence, the propagation velocity of the H-polarization is determined by

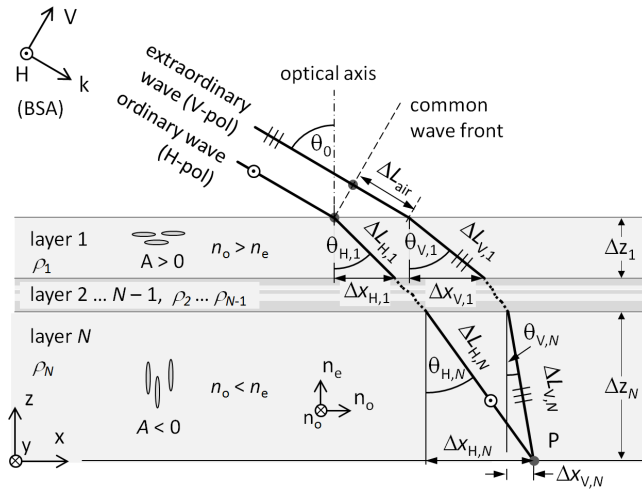


Figure 2. An electromagnetic wave (H or V polarized) is transmitted in \mathbf{k} direction with respect to the radar coordinate system (H, k, V) and with an incidence angle θ_0 with respect to the snow surface. The electric field of the H polarized wave is perpendicular to the optical axis (z), and sees the ordinary refractive index n_o (therefore called the “ordinary wave”). The electric field of the V polarized wave has a component parallel to the optical axis and is affected by the extraordinary refractive index n_e (the “extraordinary wave”). For horizontally aligned anisotropies ($A > 0$) the extraordinary wave travels faster ($n_e < n_o$) whereas for vertical anisotropies the ordinary wave is faster ($n_o < n_e$). As refraction differs for both waves, also the optical distances differ when measured from a common wave front to the same point P on the ground. Describing the radar reflection at a common point P seems arbitrarily chosen, however only reflections from common scatterers within one radar resolution cell contribute to the CPD as described in Sect. 3.2. The structural anisotropy of the layers is shown as expected for fresh snow (layer 1) deposited on top of old snow (layer N). The layer of fresh snow with density ρ_1 and thickness Δz_1 is drawn with horizontal structures with a anisotropy $A > 0$. The thick layer of old snow is drawn as vertical ice grains ($A < 0$) grown by temperature gradient metamorphism. The theory in this paper is true for any random layering of densities and anisotropies due to Snell’s law as long as absorption or volume scattering are negligible.

the ordinary refractive index n_o . The vertical polarization (V) is defined perpendicular to H and the propagation vector \mathbf{k} . The V-polarization is not parallel to the optical axis z as for side-looking radar systems the incidence angle θ_0 can never reach 90° . Therefore, the electric field of the V-polarization always has one component along the optical axis z and one component perpendicular to it, along x . For the V polarization, the refractive index n_V depends on the propagation angle θ_V in the medium and can be described by the refractive index ellipsoid (Saleh and Teich, 1991)

$$\frac{1}{n_V^2(\theta_V)} = \frac{\cos^2 \theta_V}{n_o^2} + \frac{\sin^2 \theta_V}{n_e^2}. \quad (13)$$

The refractive indices for the H and V polarized wave are²

$$n_H = n_o \quad (14a)$$

$$n_V(\theta_V) = \frac{n_o n_e}{\sqrt{n_e^2 \cos^2 \theta_V + n_o^2 \sin^2 \theta_V}}. \quad (14b)$$

The refraction at the air-snow interface is described by Snell’s law which for the H polarization is

$$n_{\text{air}} \sin \theta_0 = n_H \sin \theta_H. \quad (15a)$$

For the V polarization, the refractive index n_V depends on θ_V , which in turn depends on n_V . The modified Snell’s law

$$n_{\text{air}} \sin \theta_0 = n_V(\theta_V) \sin \theta_V \quad (15b)$$

has therefore to be solved simultaneously with Eq. (14b). It follows that

$$n_V(\theta_V) = \sqrt{n_o^2 + \left(1 - \frac{n_o^2}{n_e^2}\right) n_{\text{air}}^2 \sin^2 \theta_0}. \quad (16)$$

Equation (16) can be used in Eq. (15b) to calculate the angle θ_V . Note, that θ_V is only implicitly contained in Eq. (16) by θ_0 and Snell’s law (15b). For a birefringent medium, θ_V does no longer describe the direction of propagation of an optical beam (which does the Poynting-Vector), but instead the direction which is perpendicular to the wave fronts (the wave vector \mathbf{k}). As we are interested in the retardation of wave fronts, we use θ_V which determines the direction of \mathbf{k} in the birefringent medium. For multi-layer systems comprising N anisotropic layers which all have the optical axis parallel to the z axis, Eqs. (15a) and (15b) are valid for every layer because Snell’s law holds at each layer-interface

$$n_j \sin \theta_j = n_{j+1} \sin \theta_{j+1} \quad \text{for } j = 0, 1, \dots, N-1 \quad (17)$$

and with $n_0 = n_{\text{air}} \approx 1$.

The difference in propagation delay between both polarizations can now be calculated. Fig. 2 shows the geometry of a multilayer system where each layer j of thickness Δz_j can have a different anisotropy A_j and density ρ_j . The layers are numbered from top (1) to bottom (N). Two perpendicular polarized plane waves of frequency $\nu = \omega/(2\pi)$ described by $\mathbf{E}(t, \mathbf{r}) = \mathbf{E}_0 e^{i(\omega t - \mathbf{k} \cdot \mathbf{r})}$ are transmitted to the snow surface with an incidence angle θ_0 . For a fixed time t , the accumulated phase along a distance r is given by $\phi = \mathbf{k} \cdot \mathbf{r}$, where the magnitude $k = |\mathbf{k}|$ of the ordinary (H) and extraordinary (V) wave vectors depend on the vacuum wavelength λ_0 and the corresponding refractive indices:

$$k_H = \frac{2\pi n_H}{\lambda_0} \quad \text{and} \quad k_V = \frac{2\pi n_V}{\lambda_0}. \quad (18)$$

²Note that the equation for n_V^2 in (Leinss et al., 2014b) is an approximation of Eq. (13) for small anisotropies. The approximation follows from Eq. (13) by writing $n_o^2 = \epsilon_{\text{eff}} - \delta$ and $n_e^2 = \epsilon_{\text{eff}} + \delta$ and applying a first order Taylor expansion in δ , neglecting terms $\mathcal{O}(\delta^2/\epsilon_{\text{eff}}^2)$.

The two paths for the ordinary and extraordinary waves which connect a common wave front with a point P at the snow-soil interface are drawn in Fig. 2. The two-way phase difference between the two paths is given by

$$\begin{aligned} \phi'_{\text{CPD}} &= 2 \cdot \sum_{j=1}^N (k_{V,j} \Delta L_{V,j} - k_{H,j} \Delta L_{H,j}) + 2k_0 \Delta L_{\text{air}} \quad (19) \\ &= 2 \sum_{j=1}^N \frac{k_{V,j} \Delta z_j}{\cos \theta_{V,j}} - 2 \sum_{j=1}^N \frac{k_{H,j} \Delta z_j}{\cos \theta_{H,j}} + 2\phi_{\text{air}}. \quad (20) \end{aligned}$$

The in-air phase difference $\phi_{\text{air}} = k_0 \Delta L_{\text{air}}$ depends on the sum of horizontal displacements $\sum \Delta x_{H,j} - \Delta x_{V,j}$ and the wave vector in air, $k_0 = 2\pi n_{\text{air}}/\lambda_0$, and is given by

$$\phi_{\text{air}} = -k_0 \cdot \sin \theta_0 \sum_{j=1}^N \Delta z_j (\tan \theta_{V,j} - \tan \theta_{H,j}). \quad (21)$$

Equation (20) can be rearranged and combined with Eqs. (21) and (18) to formulate the CPD in the BSA convention (11) as

$$\phi_{\text{CPD}} = (-1) \frac{4\pi}{\lambda_0} \sum_{j=1}^N \Delta z_j \cdot \Delta \zeta(\rho_j, A_j, \theta_0). \quad (22)$$

The contributions of individual layers of thickness Δz are given by the specific path length difference

$$\Delta \zeta(\rho, A, \theta_0) = \sqrt{n_V^2 - \sin^2 \theta_0} - \sqrt{n_H^2 - \sin^2 \theta_0}. \quad (23)$$

The specific path length difference defines the optical path length difference per thickness ΔZ of an anisotropic medium observed under a surface incidence angle θ_0 . The refractive indices n_V and n_H are defined for each individual layer by Eqs. (14a) and (16) using the effective permittivity from Eqs. (12a) and (12b) derived in Sect. 2.2 for given snow density ρ and structural anisotropy A .

The horizontal structures in fresh snow cause a faster propagation speed for the VV polarization than for HH. Consequently, HH will have a larger phase delay than VV at the receiving antenna. This results in a positive CPD, $\phi_{\text{CPD}} = \phi_{\text{VV}} - \phi_{\text{HH}}$, due to the sign-change because of the BSA.

The specific path length difference, $\Delta \zeta$, increases with incidence angle (Fig. 3, left) and with increasing densities below 0.2 g cm^{-3} (Fig. 3, right). When the snow density increases beyond 0.3 g cm^{-3} , refraction reduces the alignment of the V polarization with respect to the optical axis and consequently $\Delta \zeta$ decreases (Fig. 3, right). Therefore, a broad maximum of $\Delta \zeta$ is observed for densities between 0.2 and 0.4 g cm^{-3} (Fig. 3, right), where only a weak density dependence exists.

Also, above a density of about $\rho = 0.55$, the dielectric anisotropy $\Delta \varepsilon = \varepsilon_x - \varepsilon_z$ decreases (Fig. 1, right) such that $\Delta \zeta$ vanishes at $\rho = \rho_{\text{ice}}$ where no air inclusions are present anymore. We note here, that $\Delta \zeta$ vanishes only for dielectric

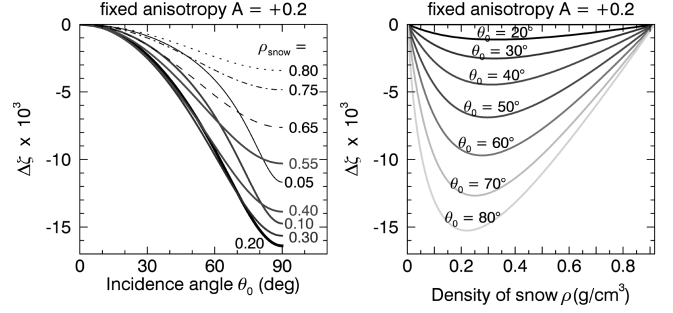


Figure 3. The specific path length difference $\Delta \zeta$, expected between vertically and horizontally polarized microwaves according to (Eq. 23), is plotted for snow with horizontally aligned oblate ice grains ($A = +0.2$) over incidence angle (left), and snow density (right).

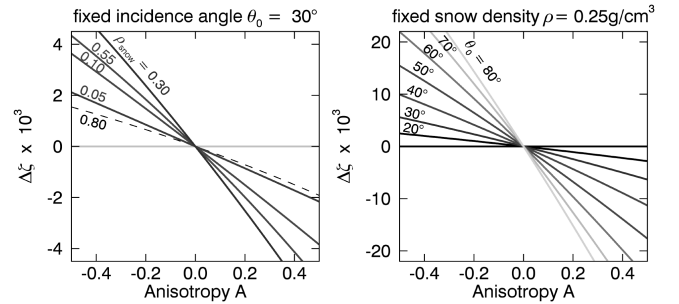


Figure 4. The specific path length difference $\Delta \zeta$, according to (Eq. 23), plotted over anisotropy for different snow densities but fixed incidence angle, $\theta_0 = 30^\circ$, (left) and for different incidence angles but fixed snow density, $\rho = 0.25 \text{ g cm}^{-3}$, (right).

isotropic (polycrystalline) ice. This is not generally the case as for ice on glaciers and ice sheets the crystal axis of ice (c-axis) can have a preferential orientation (e.g. Matsuoka et al., 1997; Fujita et al., 2014, and also Appendix A).

The weak dependence of $\Delta \zeta$ on snow density, at least for the density range of seasonal snow, allows for a quite rough estimate for snow density when the CPD is used to determine the anisotropy of snow. For seasonal snow, densities of 0.15 and 0.4 g cm^{-3} have been reported (Bormann et al., 2013). Within this range, the CPD varies by less than 20 % as shown by Fig. 3(left).

In contrast to the weak density dependence, $\Delta \zeta$ depends almost linearly on anisotropy A for all densities and incidence angles (Fig. 4, left and right). Therefore, the CPD is mainly determined by snow depth Δz and the anisotropy A which makes determination of A almost independent on snow density.

3.4 Discussion of the CPD regarding literature results

For firm (with $\rho = 0.4 \text{ g cm}^{-3}$, $\Delta \varepsilon = -0.05$), as observed by Fujita et al. (2014) for the upper 5 meters of the ice sheet at NEEM in the northwest of Greenland, we would expect

a negative anisotropy $A = -0.25$ ($A' = 1.3$) due to vertical structures. Similar firm conditions ($\rho = 0.4 \text{ g cm}^{-3}$, $\Delta\varepsilon = -0.05$) have been found by Fujita et al. (2009) at Dome Fuji in Antarctica who determined a slightly lower structural anisotropy ($A' \approx 1.15$) by means of X-ray microtomography. Similar anisotropy values of $A' = 1.2$ or greater have been observed in Antarctic firm by Alley (1987). In a recent firm core study from Dome Fuji, measurements of dielectric anisotropies $\Delta\varepsilon = -0.01 \dots -0.05$ were used as a surrogate for a geometrically anisotropic microstructure (Fujita et al., 2016). The density and dielectric measurements of the studies of Fujita indicate that a CPD of $\phi_{\text{CPD}} = -80^\circ$ per meter would have been measured for the radar parameters of the satellite TerraSAR-X as used in the following study about seasonal snow: in Leinss et al. (2014b) a CPD of $60\text{--}150^\circ/\text{m}$ was measured for fresh snow ($\rho = 0.2$) in Finland at $\theta_0 = 32.7^\circ$ and 9.65 GHz, which would correspond to elongated horizontal structures with an anisotropy between $A = +0.2$ and $+0.5$ ($A'^{-1} = 1.2$ and 1.7). Somewhat lower anisotropy values were found for natural, undisturbed and also for sieved seasonal snow where in both cases initially horizontal features ($A'^{-1} = 1.12$ and 1.17) decayed within 6 days after which in one case a preferentially vertical orientation ($A' = 1.12$) was found (Schneebeli and Sokratov, 2004).

Compared to the simplified model of Leinss et al. (2014b), where refraction was not included, we get about 5–10% lower values for the CPD using Eq. (22). The steeper propagation angle due to refraction leads to a decreasing z -component of the V-polarized field, consequently the birefringence effect is reduced. Using the weighted average of the two Hashin–Shtrikman bounds to calculate ε_{eff} leads to an additional decrease of up to 30% for higher snow densities compared to the model published by Leinss et al. (2014b).

3.5 Generalization for scattering multilayer systems

Equation (22) is valid for a multi-layer system, where scattering and absorption are negligible in or between different snow layers. In the present work, we solely concentrate on non-scattering and non-absorptive media for which all scattered energy returns from the bottom of the multi-layer snow system. For multi-layer systems where scattering occurs at the snow surface, at layer boundaries or within snow layers, or where microwave-absorbing layers are present, the location of the main scattering center is difficult to define and depends strongly on the scattering and absorption properties of the snowpack.

The scattering properties are given by the ratio of grain size to wavelength but also by the surface roughness and the dielectric contrast between neighboring layers. Scattering within the snowpack can occur e.g. in old metamorphic snow like depth hoar, in snow which contains ice layers and melt-crusts but also in deep snow on glaciers where the snow depth exceeds the penetration depth of microwaves.

In the following we briefly outline how Eq. (22) can be generalized to estimate the CPD when scattering of different layers needs to be included. In order to generalize our model for media where volume scattering cannot be neglected, we define – possibly complex – amplitude scattering factors μ_j for each layer boundary. The scattering contribution of the first layer boundary, the air/snow interface, is given by μ_0 . The phasor $e^{i\phi_1}$ defining the CPD of the first layer contributes with the backscatter amplitude factor μ_1 of the first-to-second layer boundary to the total phase difference. The reflection after the second layer accumulates the CPD of the first and of the second layer, so that the second phasor is given by $e^{i(\phi_1+\phi_2)}$ and so on. The total phase difference is then

$$\begin{aligned} \phi_{\text{CPD}} &= \mu_0 + \mu_1 \cdot e^{i\phi_1} + \mu_2 \cdot e^{i(\phi_1+\phi_2)} + \dots \\ &= \sum_{j=0}^N \mu_j \prod_{k=0}^j e^{i\phi_k} \quad \text{with } \phi_0 = 0. \end{aligned} \quad (24)$$

Scattering within layers can be accounted for by subdividing a homogeneous layer into a sufficient number of finite layers.

For homogeneously scattering and/or absorbing volumes, $|\mu_j|$ would decrease exponentially, whereas μ_j can be quite heterogeneous for ice layers which occur e.g. in the percolation zone of glaciers (Parrella et al., 2015) or for snowpacks which contain e.g. melt crusts and ice layers. In such cases, assumptions must be made for the penetration depth or the penetration depth must be determined independently and the inversion of the CPD to determine the anisotropy can quickly be questionable.

3.6 Contribution of a rough ground surface

The method to determine the anisotropy of snow as presented in the previous sections relies on the assumption that the CPD of the underlying ground is zero or at least known (see Sect. 5.6). Radar experiments have shown that the CPD is close to zero for soil at small incidence angles but shows an increasing standard deviation for rough surfaces. It has also been found that the CPD increases to a few ten degrees with incidence angle for rough soil (Sarabandi, 1992; Oh et al., 2002). The CPD is also influenced by vegetation cover, especially for oriented vegetation (Ulaby et al., 1987). Therefore, CPD measurements of snow free ground are recommended to verify if any CPD bias exists.

4 Experimental data

For the validation of our model we analyzed radar data acquired within the Nordic Snow Radar Experiment (NoSREx) campaigns (Lemmetyinen et al., 2013). The NoSREx campaigns consisted of extensive field measurements and various active and passive microwave measurements acquired at a test site near the town of Sodankylä in northern Finland.



Figure 5. The radar data and meteorological measurements were acquired on the above shown test site nearby the town of Sodankylä, Finland. The SnowScat Instrument SSI (inset) was mounted on a 9 m high tower. The reference target (sphere) used for calibration can be located behind a tree. The SSI scanned sector 1 and 2 with different azimuth (az) and incidence angles θ_0 . The meteorological sensors SMT, MétéoMast, SDAT1, GWI, and AWS are specified in Table 1. Snow profiles were taken at CT-1 to CT-4 for analysis by computer tomography. Snow density was measured in the snow pit, with the Gamma water Instrument (GWI), and was also derived from the snow water equivalent (SWE) determined by the SSI as described in Leinss et al. (2015). The variability of snow depth was measured with seven sticks inside the ellipse "SDvar".

The test site, shown in Fig. 5, is an almost flat forest clearing which is surrounded by boreal forest. The test site is covered by low taiga-type vegetation which grows on mineral soil. The test site is equipped with various sensors to measure meteorological data and snow properties. For the radar measurements, two sectors were defined on the test site, sector 1 in the center of the forest clearing, sector 2 between trees.

4.1 Microwave measurements

The radar data were acquired by the SnowScat Instrument (SSI), which was installed on a 9 m high tower. The tower is shown in Fig. 5, the SSI with its two horn-antennas is shown in the inset.

SnowScat is a fully polarimetric, coherent, continuous-wave stepped-frequency, real aperture radar and operates between 9.2 and 17.8 GHz (Wiesmann et al., 2008; Werner et al., 2010; Wiesmann and Werner, 2010). It was originally developed and built for snow backscatter measurements within the ESA ESTEC project KuScat, contract No. 42000 20716/07/NL/EL. Both antennas of the instrument can transmit and receive in horizontal (H) and vertical (V) polarization. The (-3dB) beam width of the horn antennas ranges from 5 to 12 degree, depending on polarization and frequency.

Table 1. Abbreviations for different sensors and measurements

SSI	Snow Scat Instrument (ground based radar)
sphere	Calibration target for the SSI
CT-#	Snow profile number # acquired in the field and analyzed by computer tomography (μ CT)
CT-1	Profile sampling date: 03-03-2011
CT-2	Profile sampling date: 21-12-2011
CT-3	Profile sampling date: 01-03-2012
CT-4	Profile sampling date: 28-02-2013
SDAT1	Snow Depth and Air Temperature sensor no. 1
SMT	Sensors for Soil Moisture and soil Temperature
MeteoMast	Meteorological mast (snow depth, snow temperature profiles)
SDvar	Snow Depth variability course
AWS	Automatic Weather Station
GWI	Gamma Water Instrument (SWE measurement by gamma ray absorption)

Radar acquisitions were acquired of both sectors of the test site every four hours. The sectors were scanned in azimuth-subsectors of 6° by rotating the antennas around the vertical axis (az). The scan was done for each of the four nominal incidence angles ($\theta_0 = 30^\circ, 40^\circ, 50^\circ$, and 60°) resulting in 17×4 acquisitions for sector 1 and 5×4 acquisitions for sector 2. Each subsector was measured in all four polarization combinations, VV, HH, VH, and HV, using the full frequency range. A detailed description showing the acquisition geometry, the antenna patterns and the polarimetric backscatter signal of both sectors can be found in Leinss et al. (2015); In this publication, the snow water equivalent (SWE), later used to estimate the snow density, was determined from SSI data by means of differential radar interferometry.

In the present work, it is assumed that during dry snow conditions all energy is backscattered from the ground below the snowpack. This assumption is justified because SWE could be determined with high precision in Leinss et al. (2015), where a requirement for successful SWE measurements was a snowpack transparent for microwaves. The assumption of a transparent snowpack is further supported by the analysis of the radargrams (Fig. 9 and 10 in Leinss et al., 2014a) which show a distinct range shift of 1.2 m of the antenna pattern at the onset of snow melt. This shift measured at $\theta_0 = 40^\circ$ is proportional to the slant range difference between the wet snow surface of 80 cm height, as observed in April 2012 and 2013, and the underlying ground which had been visible during the dry snow conditions before snow melt.

4.2 Meteorological instruments, snow depth and density determination

Several automated meteorological sensors were installed on the test site at the locations shown in Fig. 5. Abbreviations for the sensors are listed in Table 1; technical details can be found in Kontu et al. (2011). Snow depth and air tempera-

ture were measured by the sensor SDTA1. Soil temperature (2 sensors at 2 cm depth) and soil moisture (4 sensors, each 2 at 2 cm and 10 cm depth) were measured by a sensor network named SMT. Snow depth and temperature profiles within the snowpack were measured by the meteorological mast (MeteoMast) 100 m east of the SSI. An automatic weather station (AWS), located 500 m north of the SSI measured snow depth, air temperature, and other meteorological parameters.

The variability of snow depth on the test site was measured with seven sticks located 1 m apart, the so called "Snow depth variability course" (SDvar). The seven sticks are located in the ellipse "SDvar" in Fig. 5. Snow depth measurements done at the seven sticks show a quite homogeneous snow depth distribution with a standard deviation of 2–3 cm during dry snow conditions. This allows for comparison of snow data measured at different locations within the test site.

The depth-averaged snow density ρ_{avg} was manually measured in the snow pit once every week. ρ_{avg} was also calculated from snow depth measured by SDTA1 and from SWE measurements, as SWE in mm water column equivalent = snow depth $\cdot \rho_{\text{avg}} / \rho_{\text{ice}}$. SWE was obtained from the SSI during dry snow conditions (Leinss et al., 2015), and from measurements of gamma ray absorption within the snowpack using the Gamma Water Instrument, GWI, during wet snow conditions. A short description of the GWI can be found in Kontu et al. (2011) and Leinss et al. (2015).

4.3 Computer tomography profiles

The microstructure of four vertical snow profiles, CT-1...CT-4, sampled in the field on the dates given in Table 1, was determined using computer tomography. The location of the profiles are shown in Fig. 5. For each profile, vertically overlapping samples of about 10 cm height were taken to cover entire snow depth profiles. The samples were later analyzed by computer tomography at the WSL Institute for Snow and Avalanche Research SLF in Switzerland. For analysis by means of μCT , the snow samples had to be cast for transportation from Finland to the cold lab at SLF.

An analysis of the μCT data, which we used here to determine the anisotropy, was already published with respect to other snow structure parameters in Proksch et al. (2015). Here we briefly summarize the methodology of the casting and processing procedure.

The snow samples were cast using Diethyl-Phthalate (DEP) to preserve the snow structure. The casting procedure as well as an accuracy analysis of cast and not-cast samples are described in Heggli et al. (2009). In the cold lab, the samples were scanned with a nominal resolution (voxel size) ranging from 10 μm for new snow to 20 μm for depth hoar. The size of the evaluated volumes ranged from 67 mm^3 for CT-1, CT-2 and CT-3 ($512 \times 512 \times 256$ voxel with 10 μm voxelsize) to 917 mm^3 for CT-4 ($512 \times 512 \times 600$ voxel with 18 μm voxel size).

The 3-D-gray-scale images, which resulted from the scans, were filtered using a Gaussian filter (sigma = 1 voxel, filter kernel support = 2 voxel). The smoothed images were then segmented into binary images. For snow/air segmentation, the intensity threshold was chosen at the minimum between the DEP peak and the air peak in the histograms of the gray-scale images.

4.4 SnowScat data processing and CPD calibration

The raw data measured by the SSI in the frequency-domain for each azimuth direction az and incidence angle θ were windowed to select a specific frequency band of 2 GHz bandwidth. The selected bandwidth was then focused into range profiles of single-look-complex (SLC) format (for procession details see Leinss et al., 2015). The pixels of an SLC acquisition represent the complex-valued backscatter coefficients $S_{\text{pol}}(r, \theta_0, az)$ of each polarization $\text{pol} \in \{\text{VV}, \text{HH}, \text{VH}, \text{HV}\}$. The uncalibrated CPD for each az and θ_0 was then calculated from the copolar coherence (Eq. 9) evaluated for the measured backscatter coefficients S_{VV} and S_{HH} . The ensemble averages $\langle \cdot \rangle$ contained about 150–300 range-pixels covering the full width (-3 dB) of the antenna footprint. By summing over the antenna footprint, slightly different CPD values have been averaged due to the incidence angle variation of 5–8 degree within the common antenna footprint of both polarizations. Still, across the footprint, the incidence angle dependence of the CPD is sufficiently linear so that no systematic errors are expected. For noise and speckle reduction, the copolar coherences of different azimuth-subsectors with the same incidence angle were averaged. The phase $\phi_{\text{CPD}} = \phi_{\text{VV}} - \phi_{\text{HH}}$, obtained from the averaged coherence, is the CPD in the backscatter alignment convention as defined in Eq. (10). The CPD showed some systematic drifts, therefore a metallic sphere was used for calibration of the measurements. The calibration procedure is detailed in Appendix D.

4.5 Selecting valid acquisitions

Invalid acquisitions were removed before the analysis with the help of the calibration data. Acquisitions were classified as invalid if the CPD or the Radar Cross Section (RCS) of the reference targets (sphere, plate) deviated too far from the expected values or if the temporal trend of the sphere and the plate were not in agreement. In the two seasons before 18 November 2011, where the plate target was not installed yet, the sphere showed very stable results therefore the data was considered as valid. For sector 2, which was located between trees, some subsectors at the left and right hand side were disturbed by trees (Leinss et al., 2015, Fig. 3) and were therefore excluded from the analysis.

4.6 Measurements: Four years of CPD time series

Four years of CPD time series data, acquired by the SnowScat instrument between 2009 and 2013, are plotted in Figs. 6–9 together with meteorological measurements.

5 The upper panels of the figures show meteorological parameters (abbreviations are given in Table 1, locations are shown in Fig. 5). The snow depth measured by the three sensor SDAT1, AWS and MeteoMast are shown as dashed lines, the mean and standard deviation of the snow depth variability course "SDvar" is shown as errors bars. The blue solid line shows the average snow depth of SDAT1, AWS and MeteoMast. Air- and soil temperature of the sensor SDAT1 and SMT are plotted below the snow depth. The second panel shows soil moisture for two locations, each at 2 and 10 cm
10 depth, measured by the sensors of SMT (brown). The snow density (solid black line) was determined by dividing SWE, as described in Sect. 4.2, by the snow depth measured by the sensor SDAT1. Manual density measurements obtained in the snow pit are shown as black dots.

20 The lower panels of the figures show the polarimetric radar measurements. The CPD ($= \phi_{VV} - \phi_{HH}$) measured by the SSI is plotted for different incidence angles θ_0 (third panel), and frequencies f (fourth panel). The lowest panel shows the copolar coherence $\gamma_{VV,HH}$ for three different frequencies and the highest incidence angle $\theta_0 = 60^\circ$.

25 The dark gray shading for April and May in the figures indicates the period of snow melt. Snow free conditions are indicated by a light gray shading in autumn and May/June. In the following paragraphs we summarize the main characteristics of the measurements observed during the four winter seasons.

30 A common characteristic found in all four seasons is a rising CPD during snow fall. The CPD reached its maximum typically a few days after snowfall ended. The opposite trend, a gradually decreasing CPD during periods of cold temperatures without much fresh snow, can be observed as long as temperatures were well below 0°C . During snow melt, the CPD is close to zero as the penetration of microwaves into the wet snowpack is inhibited. Soil moisture correlates well
40 with snow melt, but does not show any influence on the CPD, even when the soil was not frozen in early winter.

45 The copolar coherence, $\gamma_{VV,HH}$, is shown for the highest incidence angle ($\theta_0 = 60^\circ$) where it is most sensitive to volume scattering. During dry snow conditions, $\gamma_{VV,HH}$ ranges from 0.4 to 0.7 with lower values for higher frequencies. Only at 16.8 GHz at 60° the coherence was found to be lower during winter (≈ 0.4) compared to snow free conditions ($\gamma_{VV,HH} \approx 0.55$, horizontal dashed line "no snow"), which indicates some weak scattering in the snow volume.
50 The highest values of $\gamma_{VV,HH} = 0.7 \dots 0.8$ were measured during snow melt, where the microwave penetration depth is very weak (a few cm) and scattering occurs at the snow surface. After all snow has melted, the coherence decreased to

$\approx 0.5 \dots 0.6$ as some volume scattering occurs at the low vegetation.

4.7 Interpretation of CPD measurements with respect to snow conditions

The four analyzed winter seasons showed quite different snow conditions. In the following, we provide an interpretation of the measured CPD time series with respect to snow properties which were observed in the field and which were documented in Lemmetyinen et al. (2013, p. 425/49).
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The winter of 2009–2010 was characterized by mild temperatures until mid of December which caused delayed freezing of the soil compared to average years. Snow accumulation happened gradually and the mild temperatures lead to larger snow densities of 0.2 g cm^{-3} in early winter compared to other years. Due to warm temperatures, depth hoar was largely absent and melt-refreeze events in early December caused the formation of a crust in the shallow snowpack which was later covered by snow. Later in winter, two major snow fall events occurred. The first happened during early February after which the CPD increased by more than 50° but decreased quickly due to strong temperature gradients causing a fast metamorphism into vertical structures. The second major snow fall occurred during the night from 2 to 3 March 2010, where a fast rise in temperatures together with 20 mm precipitation caused some snow settling. Despite additional fresh snow of low density, a slight increase in snow density can be observed in Fig. 6. The settling caused an abrupt increase of the CPD of about 20° during the night, followed by a total increase of more than 50° within the 5 following days. Snow settling and collapse of weak layers are discussed with respect to the "SnowScat anomaly on 2/3 march 2010" in Lemmetyinen et al. (2013, p.214 - 240). Our observations support their arguments, as a strong increase of the CPD is related to fresh snow, snow settling and a possible collapse of weak layers with vertical structures.
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75 The winter of 2010–2011 was characterized by very cold temperatures and a relatively thin snow cover. The strong temperature gradients lead to a distinct layer of depth hoar. The slightly negative CPD in December indicates a weak anisotropy with vertical structures in the snowpack. From January until March, the CPD increased with snow fall but was disrupted by a period of very cold temperatures in February during which the CPD decreased by 50° .
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85 The winter of 2011–2012 was characterized by initially exceptionally mild temperatures and late but intense snow fall during December. The weak temperature gradient from mid December until mid January caused almost no metamorphism into vertical structures. Therefore, a thick layer of horizontally oriented, settled fresh snow was preserved and a maximum CPD of $+135^\circ$ was observed at 7 January, 9 days after 20 cm of fresh snow. Almost no depth hoar was observed due to the insulating effect of the thick snowpack. The extremely large phase differences disappeared rel-
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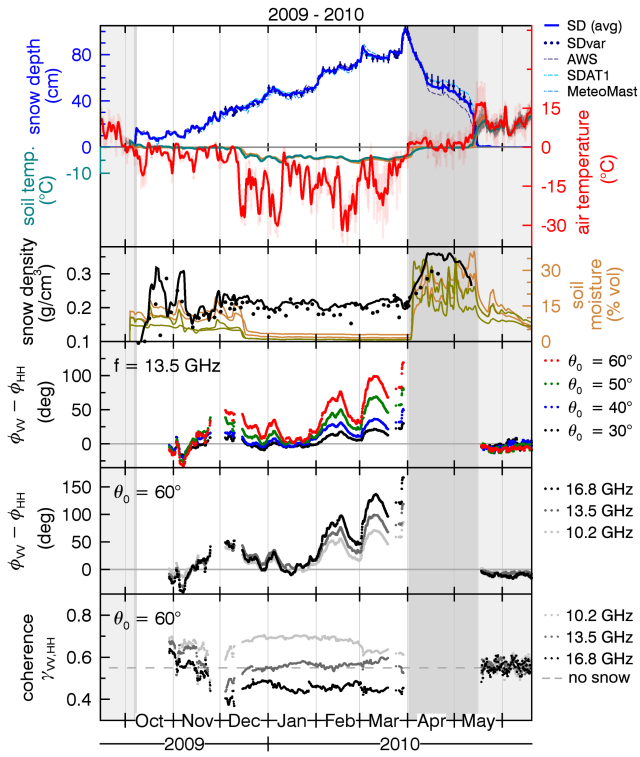


Figure 6. Winter season 2009–2010. Top panels: Meteorological data measured by the sensors described in Sect. 4.6. Bottom: CPD and copolar coherence measured by the SSI for different incidence angles and frequencies. The dark gray shading shows snow melt.

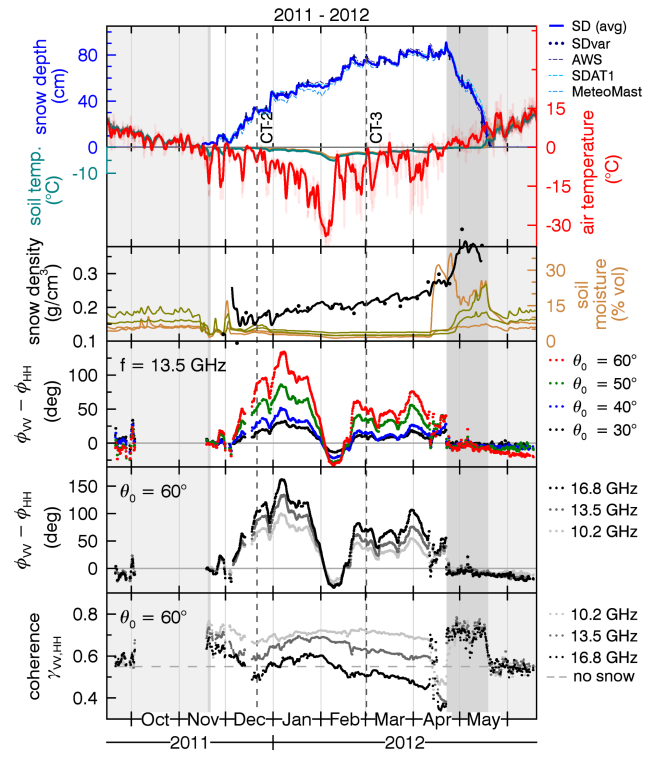


Figure 8. Winter season 2011–2012. Meteorological and radar data as shown in Fig. 6 and described in Sect. 4.6. Two vertical dashed lines shows the date when the profiles CT-2 and CT-3 were acquired.

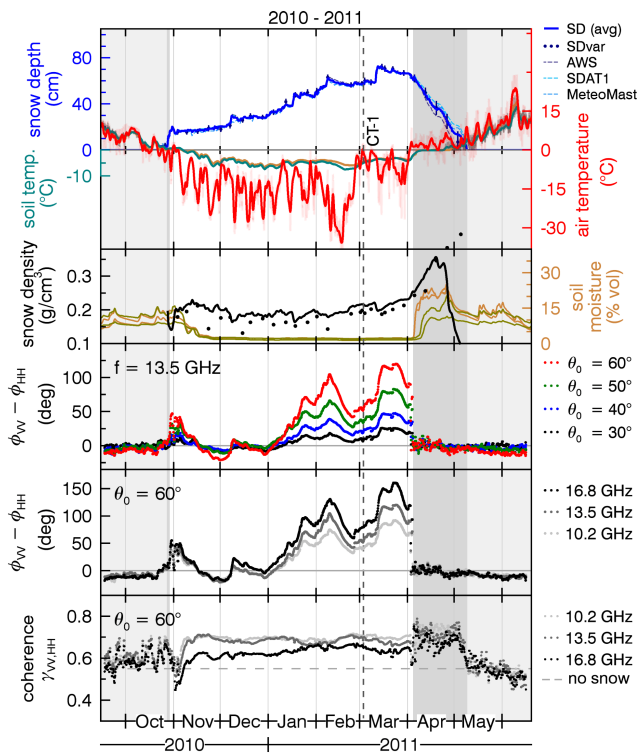


Figure 7. Winter season 2010–2011. Meteorological and radar data as shown in Fig. 6 and described in Sect. 4.6. The vertical dashed line shows the date when the profile CT-1 was acquired.

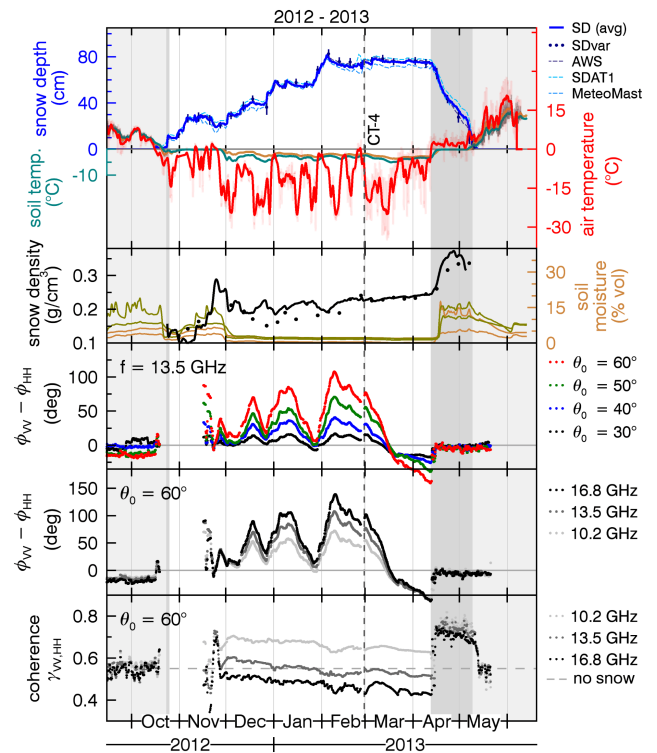


Figure 9. Winter season 2012–2013. Meteorological and radar data as shown in Fig. 6 and described in Sect. 4.6. The vertical dashed line shows the date when the profile CT-4 was acquired.

atively quickly during very cold air temperatures between -15 and -35 °C in the second half of January until mid February where the CPD even changed sign, so that a minimum CPD of -30° was observed at 9 February. After various snowfall events, the negative phase differences disappeared. At 12 April, the snow surface melted and refroze afterwards. A significant drop in the copolar coherence below snow free values (Fig. 8) indicates increased volume scattering or even residual melt water in the snowpack. A change in the backscatter pattern observed in Leinss et al. (2014a, Fig. 9) supports the observation of increased volume scattering. During the time around 12 April, when the snow surface was wet snow, the CPD dropped for a few days to zero but recovered afterwards during a short period of negative temperatures before snow melt.

The winter of 2012–2013 was again characterized by very mild temperatures but early and heavy snowfall during November, followed by three additional major snowfall events, which caused a very clear peak-like signal in the CPD. The peaks appear a few days after snow fall ended which indicates that settling of fresh snow is responsible for an increase of the CPD. In February, after the last heavy snow fall, a CPD of more than $+100^\circ$ was reached. From March until mid April, no snow fall was present and low temperatures caused a strong metamorphism for a period of 6 weeks, after which a minimum CPD of -60° was observed. With the onset of snow melt, the CPD jumped to zero due wet snow and the resulting small microwave penetration depth.

5 Analysis

5.1 Estimation of the average anisotropy of snow

The developed electromagnetic model in Sect. 2 and 3 is free of fit-parameters. Therefore, the copolar phase difference, measured and averaged over all azimuth subsectors, $CPD_{meas.}$, can be inverted with the additional information of snow depth and a good approximation of snow density (as discussed at the end of Sect. 3.3) to get a CPD-based estimate for the depth-average anisotropy, $A_{avg}^{CPD}(t, \theta_0, f)$.

For the analysis in this section, we assumed that the snowpack consisted of a single layer with a constant anisotropy. We further assumed that the snow properties (depth, density, anisotropy and also scattering properties of the underlying soil) do not vary spatially across the test site so that we can compare measurements done with different incidence angles and with different antenna footprints. The measurements of the snow depth variability course, $SDvar$, and the careful preparation of the test site’s surface support these assumptions. The area observed by the SSI covers the center of the forest clearing such that variations of snow properties due to a proximity to trees should be negligible (this is only true for sector 1, not for sector 2 located between trees). A variable snow depth due to wind drifts is unlikely, as the test site is

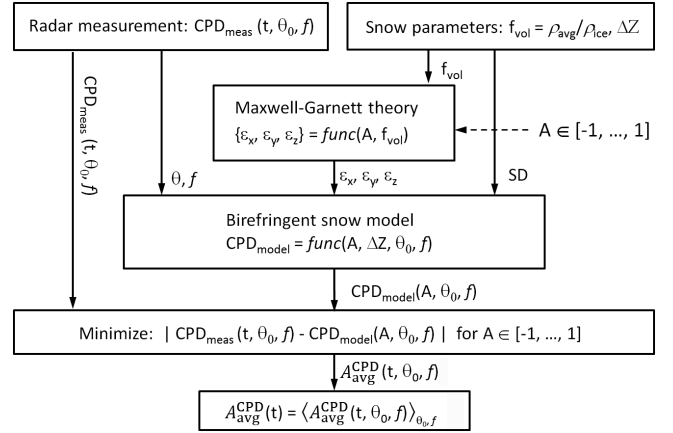


Figure 10. Processing chain used to estimate the average anisotropy of the snowpack, A_{avg}^{CPD} . The anisotropy can be estimated from the measured CPD, by minimizing the difference between modeled and measured data with respect to A if snow depth ΔZ and the ice volume fraction f_{vol} are known. The anisotropy $A_{avg}^{CPD}(t, \theta_0, f)$ was calculated independently for all incidence angles, θ_0 , and frequencies, f , and the results were averaged to obtain $A_{avg}^{CPD}(t)$.

sufficiently protected by wind due to the surrounding trees of the forest clearing (Fig. 5).

The depth-average anisotropy, A_{avg}^{CPD} , is estimated from Eq. (22) using $CPD_{meas.}$ for the radar system parameters (microwave frequency f , incidence angle θ_0). The required in-situ measured parameters snow depth ΔZ (from SDAT1), and the depth-average snow density ρ_{avg} as shown in Figs. 6 – 9. A sketch of the processing chain to determine the anisotropy is shown in the block diagram in Fig. 10. The ice volume fraction $f_{vol} = \rho_{avg}/\rho_{ice}$ follows from snow density. For every measurement time t , the depth-averaged, CPD-based estimate $A_{avg}^{CPD}(t, \theta_0, f)$ follows by minimization of the difference

$$|CPD_{meas}(t, \theta_0, f) - CPD_{model}(A(\theta_0, f, t), \theta_0, f)|. \quad (25)$$

with respect to $A \in [-1, 1]$. CPD time series at 16 different frequencies between 10 and 17 GHz and at four different incidence angles were evaluated in Eq. (25); a few of them are shown in the Figs. 6 – 9. Anisotropy values of all frequencies but only anisotropy values determined for the three larger incidence angles $\theta_0 = 40, 50,$ and 60° were later averaged to determine $A_{avg}^{CPD}(t)$, since the CPD measurements with the smallest incidence angle ($\theta_0 = 30^\circ$) showed the highest sensitivity to calibration errors. The estimated anisotropy, $A_{avg}^{CPD}(t)$, was therefore determined from $48 (= 16 \times 3)$ estimates $A_{avg}^{CPD}(t, \theta_0, f)$. The standard deviation for each time t is determined by the scatter of all $64 (= 16 \times 4)$ estimates $A_{avg}^{CPD}(t, \theta_0, f)$ around their average $A_{avg}^{CPD}(t)$. The average standard deviation of $\sigma_A \approx 0.005$ is well below the range of the obtained anisotropy between -0.05 and $+0.2$. The standard deviation varies with snow depth and is shown as a gray bar below the anisotropy.

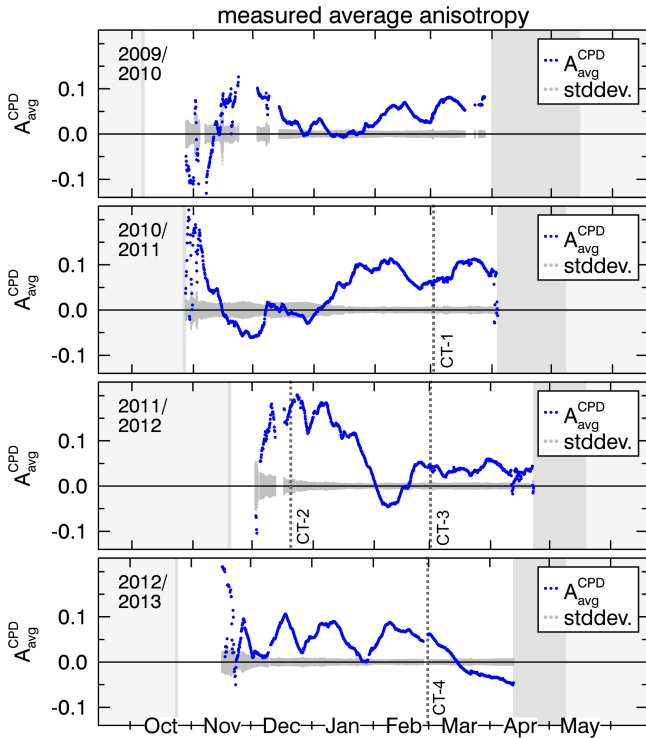


Figure 11. Average anisotropy of the snowpack, $A_{\text{avg}}^{\text{CPD}}$, determined during dry snow conditions for the winter seasons from 2009–2013. The anisotropy was derived from the CPD measured by the SnowScat instrument. The standard deviation of $A_{\text{avg}}^{\text{CPD}}$, calculated from measurements at different frequencies and incidence angles, is shown as the time-varying gray bar below the anisotropy. The dark-gray shading in April/May indicates the period of snow melt, the light-gray shadings (Oct./Nov., May/June) indicate snow free conditions. The four dashed vertical lines show the times when the anisotropy was measured by computer tomography (CT-1, -2, -3, and -4).

Time series of the estimated anisotropy $A_{\text{avg}}^{\text{CPD}}(t)$ are shown in Fig. 11. The largest positive anisotropy $A_{\text{avg}}^{\text{CPD}} \approx +0.2$ was found for Dec 2011 after intense snow fall and while temperature gradient metamorphism was very weak. The largest negative anisotropies were found for Nov 2010 ($A_{\text{avg}}^{\text{CPD}} \approx -0.06$) where strong temperature gradients in the thin snowpack were present. Large negative anisotropies were also found in Feb 2012 ($A_{\text{avg}}^{\text{CPD}} \approx -0.05$) and April 2013 ($A_{\text{avg}}^{\text{CPD}} \approx -0.05$) after periods of very cold temperatures without precipitation. As $A_{\text{avg}}^{\text{CPD}}$ is the depth-averaged anisotropy of positive and negative values much larger anisotropies are expected to be found in individual layers (see Sect.5.3 and Fig. 16).

While discussing the structural anisotropy of snow which has been derived from dielectric anisotropy it seems relevant to recall that single ice crystals also show a birefringence. Fujita et al. (2014) reported that the dielectric anisotropy due to oriented crystal fabrics is often much lower than dielectric

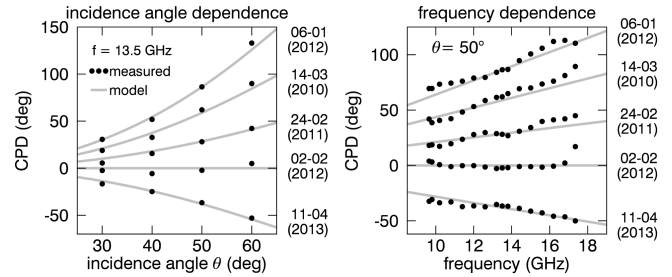


Figure 12. Left: incidence angle dependence of measured CPD vs. modeled incidence angle dependence. Right: frequency dependence of measured CPD vs. modeled linear frequency dependence.

anisotropy expected from a structural anisotropy. In appendix A we used measurements of the crystal orientation fabric in snow from Riche et al. (2013) to estimate a maximum dielectric anisotropy of $\Delta\epsilon = -0.002$ which corresponds to a structural anisotropy of $A = -0.02$. This is small compared to the measurements shown in Fig. 11 and confirms therefore the statement of Fujita et al. (2014). It is worth to note, that the dielectric anisotropy due to the vertical crystal orientation of fresh snow has the opposite sign as the dielectric anisotropy due to the horizontal structural orientation of fresh snow.

5.2 Incidence angle and frequency dependence

The larger the incidence angle, the better are the vertically polarized microwaves aligned with the optical axis of birefringent snowpack. The CPD must therefore increase with increasing incidence angle. This has already been observed in the CPD time series plotted for different incidence angles in the middle panel of Figs. 6–9.

The electromagnetic model presented in Sect. 3 predicts a nonlinear incidence angle dependence due to refraction in the snowpack (Fig. 3, left). To verify the nonlinear incidence angle dependence, we selected five dates spread over the four winter seasons to cover the maximum available range of CPDs. For each date we used the measured snow density ρ and snow depth ΔZ together with the averaged CPD-based anisotropy, $A_{\text{avg}}^{\text{CPD}}$, to model the expected incidence angle dependence. A comparison of modeled and measured incidence angle dependence is shown in Fig. 12 (left) for the five selected dates.

The CPD is modeled to be proportional to the depth of a snowpack which is transparent for microwaves. The deeper the snow and the higher the frequency, the more wavelengths “fit” in the propagation path length through the snow volume and the higher is the expected phase difference. This frequency dependence is described by Eq. (22) which shows a linear frequency dependence ($\propto \lambda^{-1}$). Larger CPD values were indeed measured for higher frequencies as it is shown for $f = 10.2, 13.5$ and 16.8 GHz in the 2nd-last panel of Figs. 6–9. For a more quantitative insight, we plotted the

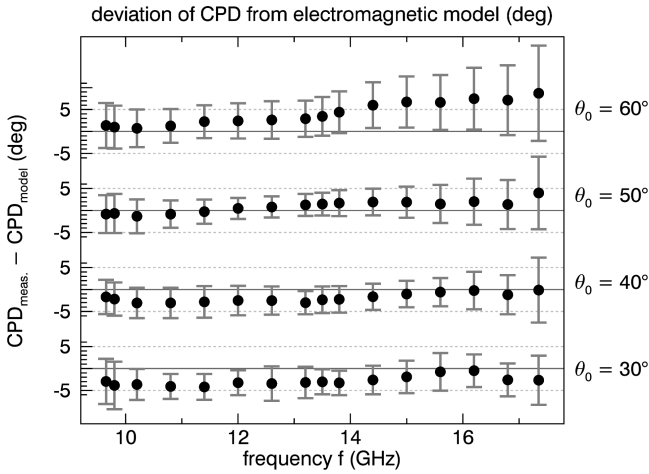


Figure 13. Deviation of measured and modeled CPD for different frequencies f and different incidence angles θ_0 . Dots show the mean deviation $\text{CPD}_{\text{meas.}} - \text{CPD}_{\text{modeled}}$ of all data acquired during dry snow conditions. The error bars are the standard-deviations calculated from about 5600 measurements.

CPD measured for 16 different frequencies in Fig. 12 (right). The CPD was plotted for the same five dates shown in Fig. 12 (left). As expected, the CPD shows approximately a linear dependence on frequency.

In order to get a better quantitative measure how well the electromagnetic model fits to measured data, we did a statistical analysis and compared the modeled phase difference, $\text{CPD}_{\text{model}}(A_{\text{avg}}^{\text{CPD}}(t), \theta_0, f)$, according to Eq. (22), with the measured phase difference, $\text{CPD}_{\text{meas.}}(t, \theta_0, f)$. The mean deviation, as well as the standard deviation of $\text{CPD}_{\text{model}} - \text{CPD}_{\text{meas.}}$, were calculated over all acquisitions acquired during dry snow conditions separately for each incidence angle θ_0 and for each frequency f .

The mean deviation is plotted over frequency and for each incidence angle in Fig. 13. The error bars indicate the standard deviation. The mean deviation is about $\pm 4^\circ$ (black dots in Fig. 13) and is almost always within the standard deviation (error bars). Only for $\theta = 60^\circ$ and $f > 14 \text{ GHz}$, we measure larger deviation up to $+8^\circ$. Figure 13 shows that neither large deviations from the expected incidence angle dependence nor large deviations from the linear frequency dependence were found. The deviations of $\text{CPD}_{\text{meas.}}$ from the $\text{CPD}_{\text{model}}$ are within the estimated calibration accuracy of $\pm 15^\circ$.

As measured and modeled data agree within a few degree, we conclude that our electromagnetic model is able to explain the observed CPD by considering snow as an optically anisotropic medium. The linear dependence on frequency confirms our assumption that the CPD is a volumetric property of snow.

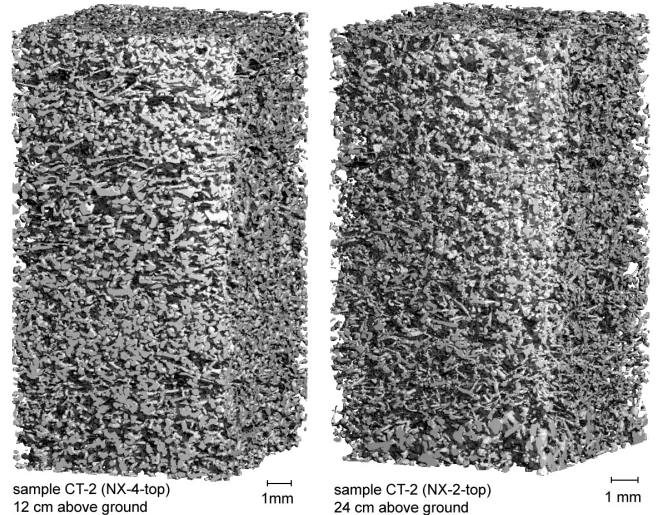


Figure 14. Two samples from the profile CT-2 (21 December 2011) taken 12 cm (left) and 24 cm (right) above ground. Horizontal structures are clearly visible in the left image but can also be identified in the right image. The CT-based average anisotropy of the two samples are $A_{\text{avg}}^{\text{CT}} = +0.26$ (left) and $+0.16$ (right). The vertically resolved anisotropy, A_{CT} , determined every 2 mm depth by means of μCT , is plotted in Fig. 16 (top right) for both samples as blue dots.

5.3 Validation with computer tomography

For validation we compared the CPD-based estimates $A_{\text{avg}}^{\text{CPD}}(t_i)$ to tomography based estimates $A_{\text{avg}}^{\text{CT}}(t_i)$ obtained from in-situ snow measurements. The four dates t_i , when the samples for computer tomography analysis were taken from the four snow pits, CT-1...CT-4, are indicated as dashed vertical lines in Figs. 7 - 9, and also in Fig. 11. Two examples of the 3-D images obtained by computer tomography are shown in Figs. 14 and 15.

In order to obtain the anisotropy from the computer tomography data, the binary 3-D images were analyzed by means of spatial correlation functions according to Löwe et al. (2011). Exponential correlation lengths, $p_{\text{ex},x}$, $p_{\text{ex},y}$, and $p_{\text{ex},z}$, were derived from the correlation functions as described by Mätzler (2002). The anisotropy determined by computer tomography, A_{CT} , is defined analogue to Eq. (1). Due to the symmetry in the x and y direction, $p_{\text{ex},x}$ and $p_{\text{ex},y}$ were averaged:

$$A_{\text{CT}} = \frac{(p_{\text{ex},x} + p_{\text{ex},y}) - 2p_{\text{ex},z}}{\left[\frac{1}{2}(p_{\text{ex},x} + p_{\text{ex},y}) + p_{\text{ex},z}\right]}. \quad (26)$$

The anisotropy was determined for the entire snow profile with a vertical resolution of 1–2 mm, depending on snow grain size. The obtained anisotropy profiles are shown in Fig. 16. For comparison, we added horizontal lines, which show the average anisotropy, $A_{\text{avg}}^{\text{CT}}$, determined from computer tomography and the average anisotropy, $A_{\text{avg}}^{\text{CPD}}$, determined from the CPD.

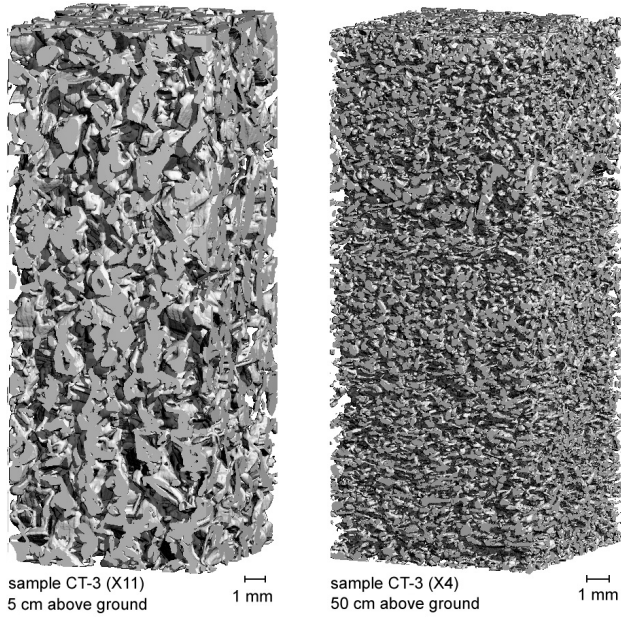


Figure 15. Two samples from the profile CT-3 (1 March 2012). The left profile, taken 5 cm above ground, shows old metamorphic snow (depth hoar) with vertical structures ($A_{\text{avg}}^{\text{CT}} = -0.24$). The profile on the right, taken 50 cm above ground, shows horizontal structures ($A_{\text{avg}}^{\text{CT}} = +0.35$) of fresh, settled snow which fell two weeks before the sample was taken. The vertically resolved anisotropy, A_{CT} , determined every 2–5 mm depth by means of μCT , are plotted in Fig. 16 (top right) for both samples as blue dots.

The first profile shown in Fig. 16 (CT-1) shows a slightly larger anisotropy, $A_{\text{avg}}^{\text{CPD}} = 0.05$, compared to the average anisotropy derived from the CT data, $A_{\text{avg}}^{\text{CT}} = 0.023$. For the profile CT-1 only a limited number of data points was available with missing data from the lowest 10 cm. However field observations shows depth hoar for the bottom 10 cm indicating that $A_{\text{avg}}^{\text{CT}}$ should even be smaller.

For the second and third profiles, CT-2 and CT-3, many CT data point were available and the difference in anisotropy is remarkably small and agrees within values of $+0.008$ and -0.004 , or $+4$ and -8% relative to the anisotropy measured by CT of $A_{\text{avg}}^{\text{CT}} = +0.16$ and $+0.05$.

For the fourth profile, CT-4, a larger difference of $+0.08$ was observed ($A_{\text{avg}}^{\text{CT}} = -0.02$, $A_{\text{avg}}^{\text{CPD}} = +0.06$). The difference might originate from a very sparse sampling of the top snow layers (Fig. 16, bottom right), as taking samples was difficult due to soft fresh snow. No samples could be taken from the top 4 cm.

For CT-4, we can exclude limited penetration as a reason for the difference, despite occurring warm temperatures a few days before, because the copolar coherence (Fig. 9) and the temporal coherence (Leinss et al., 2015, Fig. 19) did not show any anomaly. However, we can not exclude the fact, that the assumption of oriented spheroids in our model is

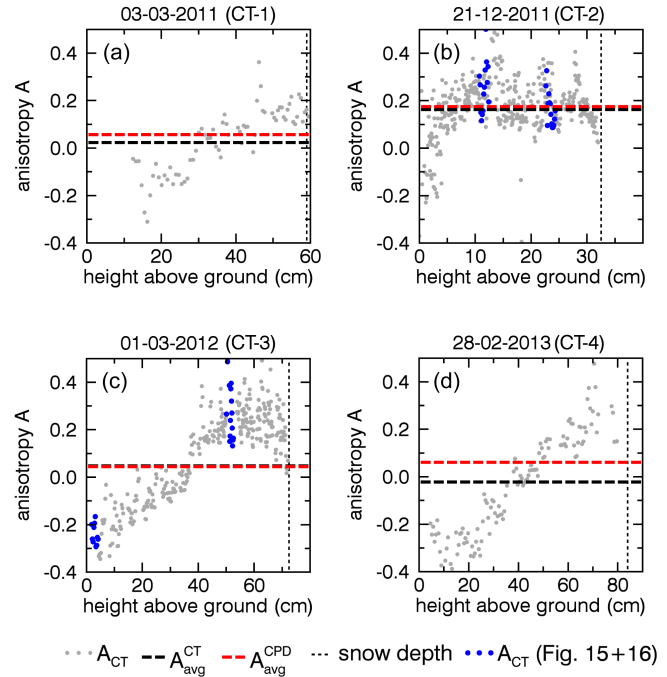


Figure 16. Vertical profiles of the anisotropy, A_{CT} , determined from computer tomography. For comparison, we plotted in each graph horizontal lines which show the depth-averaged anisotropies, $A_{\text{avg}}^{\text{CT}}$ together with the average anisotropy, $A_{\text{avg}}^{\text{CPD}}$, determined from the radar data (Fig. 11). (a): the profile CT-1 shows 30 cm of old snow with vertical structures from Nov./Dec. 2010 which is covered by metamorphic snow which fell from mid January to early February 2011 (see Fig. 7) (b): the profile CT-2 shows homogeneously distributed positive anisotropies which result from heavy snow fall during mild temperatures in December 2011 (see Fig. 8). (c): the profile CT-3 shows a thick layer with vertical structures of metamorphic snow in the lower 40 cm of the snowpack. In the upper 40 cm horizontal structures are visible which result from fresh snow fall mid of February 2012 (see Fig. 8). (d): alternating snow fall and cold temperatures lead to an almost linearly increasing anisotropy in the profile CT-4 from late February 2013 (see Fig. 9).

a too strong assumption for the very dendritic shape of fresh fluffy snow.

The vertical structure of the anisotropy profiles agrees to our expectation regarding the meteorological conditions as described in the caption of Fig. 16. In the anisotropy profiles vertical structures were found in the older snow layers, as it is expected for the geometry of metamorphic snow which was exposed to temperature gradients. In contrast to the old layers, the top layers show horizontally aligned structures as we expect it to be the case for fresh snow. The fact, that fresh snow is related to horizontal structures and therefore to a positive CPD, makes it possible to use the CPD for detection of fresh snow.

5.4 Correlation between fresh snow and a positive CPD

The settling of new snow at intermediate times can cause an increasingly positive anisotropy due to the horizontal alignment of dendrite backbones (Löwe et al., 2011). According to our theory, increasing anisotropies cause an increase of the CPD. This makes it possible to use a change in CPD to detect fresh snow as done in Leinss et al. (2014b) using satellite data.

As fresh snow settles within a few days after deposition, it is expected that the CPD does not increase simultaneously with the accumulation of fresh snow, but increases with a time-lag τ after snow fall. The parameter τ characterized the time shift between snow fall and the buildup of horizontal structures. A value $\tau \approx 0$ indicates that the CPD increases instantaneously with snow fall, whereas $\tau > 0$ gives an approximation for the time delay after which the growth of vertical structures (driven by temperature gradient) exceeds the buildup of a horizontal structures due to settling. For too large temporal offsets between a change of snow depth (SD) and a change of the CPD, no correlation is expected as temperature gradient metamorphism dominates the evolution of the CPD.

In the following, we analyze the correlation between changes in snow depth ΔSD and a change in CPD, ΔCPD . The correlation is defined as

$$R = \text{corr}\{\text{CPD}(t + \tau) - \text{CPD}(t + \tau - \Delta T), \text{SD}(t) - \text{SD}(t - \Delta T)\}, \quad (27)$$

where SD is the measured snow depth, τ the temporal offset between both time series as explained above, ΔT the sampling interval, and R is the Pearson-correlation coefficient. The sampling interval ΔT is the time difference between two measurements of snow depth and CPD. ΔT corresponds e.g. to the repeat time of satellite acquisitions. The sampling interval ΔT needs to be large enough in order to give fresh snow some time for settling such that the CPD increases above the level of phase noise. However, the sampling time should not be too large, as minor snow fall events might be missed, and also snow metamorphosis will reduce measured values of positive CPD changes as they are typical for fresh snow fall.

The scatter plot in Fig. 17 (top) shows the correlation between the depth of fresh snow within 12 days and the corresponding change in CPD measured with a time-lag of 3.5 days. The scatter plot is shown for the best correlation, $R = 0.75$, which was found for different values of ΔT and τ . The correlation coefficient R is shown for all tested values of Δt and τ in the contour plot of Fig. 17 (top right). The red cross marks the pair with the highest correlation coefficient.

The range of optimal sampling intervals, ΔT , can be derived from the contour plot shown in Fig. 17. The plot shows that the optimal ΔT is between 9 and 15 days. We analyzed all frequencies and incidence angles and the best cor-

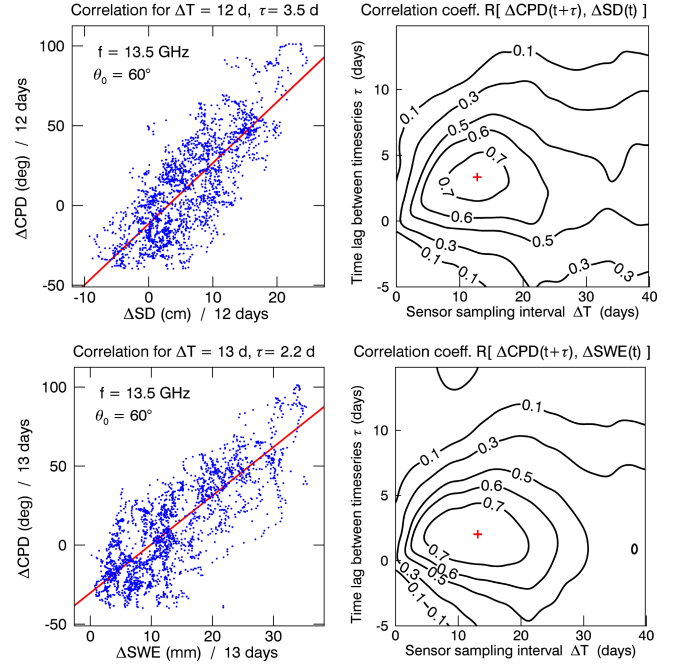


Figure 17. Left: correlation between ΔCPD and changes in snow depth ΔSD (top) and ΔSWE (bottom) within a sampling interval of $\Delta T = 12$ and 13 days. The time when the CPD difference was obtained, $t + \tau$, was shifted by $\tau = 3.5$ days (top) and $\tau = 2.2$ days (bottom) vs. the time t when the snow depth difference ΔSD was obtained because the maximum CPD was always observed when fresh snow has already settled. Right: Pearson-correlation coefficients R for different pairs of ΔT and τ shown as contour plots. The pair $(\Delta T, \tau)$ with the highest correlation coefficient is marked by a red cross.

relation coefficients, which ranged between 0.65 and 0.75, were always found for $\Delta T = 11 \pm 3$ days and a time-lag of $\tau = 3.0 \pm 0.5$ days.

The optimal sampling interval ΔT matches the 11 day orbit repeat time of TerraSAR-X. Using time series of TerraSAR-X, a CPD change of $+10$ to $+15^\circ$ per 10 cm of fresh snow was observed at 9.65 GHz at an incidence angle of 33° (Leinss et al., 2014b). From these results we would expect that the CPD changes by $40\text{--}60^\circ$ at the central frequency of the SSI of 13.5 GHz at $\theta_0 = 60^\circ$. Here we observed a change in CPD of 38° per 10 cm of fresh snow at 13.5 GHz which fits well with respect to the uncertainty $R = 0.74$ of Fig. 17 (top left).

The availability of accurate time series of the SWE measurements published in Leinss et al. (2015) made it also possible to check if a correlation exists between ΔSWE and ΔCPD . The lower two graphs of Fig. 17 show an example for the correlation. The best correlations ($R \approx 0.65\text{--}0.8$) were found for a sampling interval of $\Delta T = 10 \pm 3$ days with a time-lag of $\tau = 2.2 \pm 0.3$ days. The correlation with ΔSWE is slightly better compared to the correlation with ΔSD .

5.5 Comparison with satellite data

The CPD observed by the ground-based SnowScat instrument could also be measured from space with the satellite TerraSAR-X (TSX). Spatial and temporal correlations between the CPD and snow depth were published by Leinss et al. (2014b). Fig. 18 compares phase differences measured by TSX for the two seasons 2011–2012 and 2012–2013. The space-borne measurements show the same trends as the ground based measurements. However, the phase differences observed by TSX are about a factor 2 smaller than the CPD measured with the SSI (scatter plot in Fig. 18). The reason is very likely, that the TSX data were obtained from large open areas. In the large areas about 30 % less snow depth was measured (cf. Fig. 3 in Leinss et al., 2014b), probably due to a stronger wind exposition compared to the more wind-protected forest clearing, where the SSI was located. Wind might also be a reason for disturbed snow settling as wind drifted snow crystals show a different microstructure than undisturbed settled snow. The lower snow depth and the stronger wind exposition might explain, why smaller phase differences were measured. Some residual vegetation and trees which were contained in the large areas observed by TSX, also decreased the measured CPD due to spatial averaging.

5.6 Effect of underlying soil

Sector 2, as shown in Fig. 5, was covered with an metallic mesh by August 2011 to isolate purely snow specific radar signatures from effects of the underlying soil. In the winter 2011/12 strong ice built up on the mesh causing high backscattering. However, we did not observe any effect on the CPD and the data of both sectors agree very well (Fig. 19, middle). To prevent the build up of an ice crust in the next season, the mesh was cleared from ice on 12 December 2012 (vertical dashed line). The removal of the ice crust in the season 2012/13 did again not much affect the measured CPD, and no large differences between the soil sector and the mesh-sector were found. We could speculate, that slightly larger CPD values measured between January and April 2013 might indicate the missing of a layer of vertical oriented depth hoar crystals, but the deviation could also originate from slightly different snow conditions of the two sectors. Still, the good agreement between the measurements of the soil sector and the measurements from the metallic mesh confirms again that the measured CPD is almost purely a signal resulting from the snow volume. Although the CPD signal is caused by the snow volume, temperature gradient metamorphism alters the anisotropy of snow. As the temperature gradient is partially determined by the temperature of the underlying soil there exists an indirect effect of the soil energy balance to the evolution of the CPD.

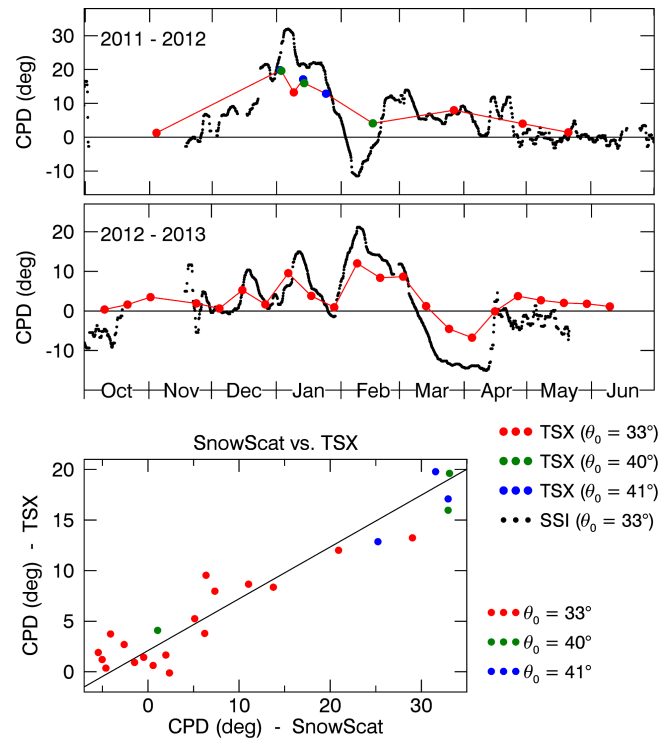


Figure 18. CPD measured by TerraSAR-X (TSX) at $\theta = 33, 40$, and 41° compared to SnowScat (SSI) measurements ($f = 9.65$ GHz, interpolated to $\theta = 33^\circ$). Both instruments show the same trend (top) but the CPD measurements of the SSI are about a factor of 2 larger than the TSX measurements (bottom). The discrepancy can be explained by different snow conditions as the TSX data were acquired over large open areas where about 30 % less snow depth was measured, compared to the test site of the SSI.

6 Conclusions

We demonstrated a contact-less technique for monitoring the temporal evolution of the depth-averaged anisotropy of a seasonal snowpack. The technique is based on measuring the birefringent dielectric properties of snow at microwave frequencies where scattering effects can be neglected. The anisotropy was determined from the copolar phase differences (CPD) measured by a ground based radar instrument complemented by the required additional data for snow depth and density.

A theoretical framework was developed, which describes the structural anisotropy of snow as oblate or prolate spheroidal ice grains with their symmetry axis in the vertical. Using Maxwell–Garnett type mixing formulas, the effective permittivity tensor was calculated to describe the microwave birefringence of snow. To make contact to the microstructure characterization in previous work, we have shown that the model of spheroidal inclusions is equivalent to a more general approach for the effective permittivity tensor based on correlation functions. From the effective permittivity tensor we calculated the birefringence and wave propagation ac-

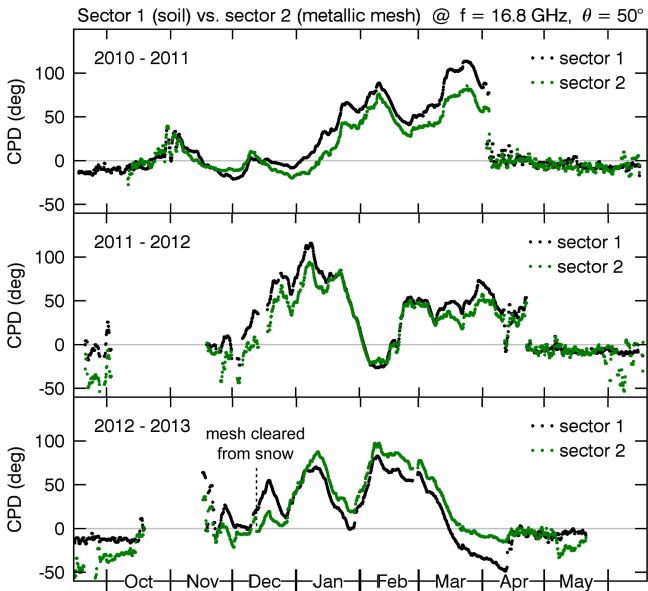


Figure 19. Comparison of the CPD measured on the two sectors of the test site. Sector 2 was located between trees behind the SSI and was covered with a metallic mesh during the last two seasons of the experiment (after August 2011). Generally, the CPD on sector 2 evolves very similar to Sector 1 and shows no large deviations.

cording to anisotropic optics. The propagation delay difference of orthogonally polarized microwaves was described in terms of the CPD. The CPD depends linearly on frequency and anisotropy but shows only a weak dependence on density for the density range of seasonal snow. The CPD was then analyzed together with the measured depth and density of the snowpack to estimate the dielectric anisotropy and to derive then the structural anisotropy averaged over the snow depth.

Four years of polarimetric radar data acquired by the SnowScat Instrument, installed at a test site near the town of Sodankylä, Finland were analyzed. Copolar phase differences ranging from -30° to $+135^\circ$ were measured for 50–60 cm of snow at a frequency of 13.5 GHz. The large variation of CPD values shows that the anisotropy of snow cannot be neglected when analyzing the CPD within polarimetric microwave studies of snow covered regions.

Overall, the depth-averaged anisotropy ranges between -0.05 and $+0.25$, with a standard deviation of 0.005 which was obtained from measurements of different incidence angles and frequencies. Additional uncertainties which originate from snow depth and density measurements were not taken into account, though.

The CPD obtained from the electromagnetic model with the anisotropy determined for each time, $A_{\text{avg}}^{\text{CPD}}(t)$, was calculated for different frequencies between 10 and 17 GHz and for different incidence angles between 30° and 60° in order to analyze deviations between modeled and measured CPD data. The modeled CPD deviated only by $5\text{--}10^\circ$ from the

measured values ranging from -30° to $+135^\circ$ and the expected linear frequency dependence could be confirmed. The linear frequency dependence confirms our assumption that the CPD is a volumetric property of snow which is determined by the dielectric anisotropy and related to the structural anisotropy of the ice matrix and pore spaces of snow.

For four dates, the CPD-based anisotropy estimates were validated by micro-computed tomography (μCT) measurements for which the anisotropy was computed directly from the two-point correlation functions. In two cases, μCT -based values for the depth averaged anisotropy agreed with their CPD based counterparts within 4 and 8%. In one case we found a fair agreement, while for the fourth sample we found a larger deviation. The origin could only be hypothesized to result from missing snow samples, limitations of the Maxwell–Garnett mixing formulas or limitations of using exponential correlation lengths to evaluate the anisotropy parameter Q .

In addition, we investigated the potential of how a changing CPD can be used to detect the accumulation of fresh snow and the increase of snow water equivalent (SWE). A weak correlation was found and an optimal acquisition interval of 8–15 days was determined to detect the depth of fresh snow from CPD measurements. It was observed that the evolution of the CPD shows a delay of about 2–3 days compared to the evolution of snow depth, which indicates an average settling time of a few days.

The CPD measurements obtained from the ground based instrument SnowScat were compared with space borne data from the radar satellite TerraSAR-X analyzed over large open areas located a few hundred meters from SnowScat. Both sensors showed the same temporal trend. However, the CPD observed by TerraSAR-X was about a factor of two smaller than the measurements done by SnowScat. The reason could be spatial variability of snow depth and snow properties due to wind exposition but also some disturbing vegetation cover in areas observed by TerraSAR-X.

Our study shows that remote sensing techniques allow determination of the dielectric anisotropy of the snowpack when the additional information about snow depth and a rough approximation of density is available. Currently, snow depth is mainly estimated from optical measurements such as photogrammetry (Marti et al., 2016; Bühler et al., 2015) or lidar instruments (Deems et al., 2013). However, the applicability of high frequency radar instruments are currently discussed in (Evans and Kruse, 2014). Snow density could potentially be derived from measurements of the snow water equivalent (Leinss et al., 2015) if data about the snow depth is available.

The possibility to observe the dielectric anisotropy of the snowpack by remote sensing techniques opens a new field of applications. Determination of the structural anisotropy and detection of fresh snow is discussed in this paper. In principle, the CPD measured over glaciers and ice sheets should provide some information about the structure of firm. How-

ever, the interpretation is difficult, though the depth of the scattering center for firn can be determined by independent means (Weber Hoen and Zebker, 2000).

Another interesting application is using CPD measurements as an indicator for the thermal conductivity of the snowpack. As the dielectric anisotropy can be exactly related to the anisotropy employed for parametrization of the thermal conductivity (Löwe et al., 2013) it seems feasible to aim at a proxy for the thermal conductivity from radar measurements, given a reasonable assumption about the mean density and snow depth. Thereby, the anisotropy would reflect predominant variations in the metamorphic state of the snowpack since increasing vertical structures are indicative of depth hoar. This might be important for the ground thermal regime in permafrost regions, if large vertical structures are created by high temperature gradients in the shallow snowpack in early-winter. Depth hoar, with its large crystals and low density close to vegetation and soil in turn, is not only important for the survival of many rodents (Bilodeau et al., 2013) but is also very important for understanding the backscattering signal from snow (King and Derksen, 2015).

The large observation time spanning four winter seasons with a sampling interval of four hours builds a unique data set to study the evolution of snow anisotropy to gain further insight into the growth mechanisms of anisotropic snow crystals. Understanding the structural anisotropy of snow enhances the understanding of macroscopic anisotropic properties such as thermal conductivity, mechanical stability and electromagnetic properties, especially the dielectric anisotropy. The developed method to measure snow anisotropy, its good agreement with ground-based μ CT measurements, and the fair agreement with satellite-based radar measurement, provide a unique opportunity to improve snow models, and globally sense the metamorphic state of the snowpack.

Appendix A: The fabric anisotropy of snow due to crystal orientation

For radio and microwaves, single crystals of hexagonal ice are dielectrically anisotropic, since the dielectric permittivity parallel to the c -axis, ε_{\parallel} , is by about $\Delta\varepsilon_{\text{ice}} = \varepsilon_{\parallel} - \varepsilon_{\perp} = 0.03 \dots 0.04$ larger compared to perpendicular permittivity ε_{\perp} (Fujita et al., 1993; Matsuoka et al., 1997). Ice and snow typically occur as polycrystals which can be characterized by their crystal orientation fabric, i.e. the distribution of c -axes. The crystal orientation fabric can be anisotropic as well which is described by an orientation tensor $\mathbf{a}^{(2)}$. This second order tensor is explained in (Durand et al., 2006), nicely visualized in (Woodcock, 1977), and was used in (Riche et al., 2013) to characterize the crystals orientation in seasonal snow. The eigenvalues of $\mathbf{a}^{(2)}$ follow the relation $1 \geq \lambda_1 \geq \lambda_2 \geq \lambda_3 \geq 0$ with $\lambda_1 + \lambda_2 + \lambda_3 = 1$ (Durand et al., 2006) and give the axis of an ellipsoid.

For snow, where the symmetry axis is the vertical (z) the eigenvalues correspond to the axis of an ellipsoid aligned along the coordinate axis $i = \{x, y, z\}$. The ellipsoid can have the following three shapes: 1) for a preferential orientation of the c -axis in the horizontal plane (horizontal girdle) the eigenvalues ($0 \leq \lambda_z < 1/3 < \lambda_y \approx \lambda_x \leq 1/2$) form an oblate spheroid. 2) for an isotropic distribution, the eigenvalues form a sphere ($\lambda_x = \lambda_y = \lambda_z = 1/3$), and 3) when the c -axis clusters around the vertical (single maximum fabric), the eigenvalues form a prolate spheroid ($0 \leq \lambda_x \approx \lambda_y < 1/3 < \lambda_z \leq 1$).

The effective permittivity of snow, ε_{eff} , composed of air pores and a matrix of (isotropically) oriented ice crystals can be described with mixing formulas (e.g. Maxwell-Garnett as in Sect. 2.2). In the following, we parametrize the effective permittivities $\varepsilon_{i,\text{eff-oi}}$ of snow comprising oriented ice crystals by a weighted average of the two permittivities of ice, ε_{\parallel} and ε_{\perp} . The weighting is determined by the eigenvalues of the orientation tensor. The weight for ε_{\parallel} is equivalent with the eigenvalue λ_i , and the weight for ε_{\perp} in the perpendicular direction follows by $(1 - \lambda_i)$. The effective permittivities for the x , y , and z -direction follow as

$$\varepsilon_{i,\text{eff-oi}} = f(\rho) \cdot [\lambda_i \cdot \varepsilon_{\parallel} + (1 - \lambda_i) \cdot \varepsilon_{\perp}]. \quad (\text{A1})$$

The function $f(\rho) = \varepsilon_{\text{eff}}(f_{\text{vol}}, \varepsilon_{\text{air}}, \varepsilon_{\text{ice}}) / \varepsilon_{\text{ice}}$ accounts for the nonlinear density dependence of the permittivity parametrized by the ice volume fraction f_{vol} and by the dielectric constant of polycrystalline ice, ε_{ice} (cf. Eq. 3).

The dielectric anisotropy of snow due to oriented ice crystals, $\Delta\varepsilon_{\text{snow-oi}} = \varepsilon_{x,\text{eff-oi}} - \varepsilon_{z,\text{eff-oi}}$, can now be related with the dielectric anisotropy of ice $\Delta\varepsilon_{\text{ice}}$:

$$\Delta\varepsilon_{\text{snow-oi}} = \varepsilon_{x,\text{eff-oi}} - \varepsilon_{z,\text{eff-oi}} = f(\rho) \cdot (\lambda_x - \lambda_z) \Delta\varepsilon_{\text{ice}} \quad (\text{A2})$$

For the two extreme cases of snow with completely vertically oriented ice crystals ($\lambda_{x,y} = 0, \lambda_z = 1$) follows $\Delta\varepsilon_{\text{snow-oi}} = -f(\rho) \Delta\varepsilon_{\text{ice}}$; for snow with a uniform orientation in the horizontal plane ($\lambda_{x,y} = 0.5, \lambda_z = 0$) one obtains $\Delta\varepsilon_{\text{snow-oi}} = +\frac{1}{2} f(\rho) \Delta\varepsilon_{\text{ice}}$. For the isotropic case ($\lambda_i = 1/3$) one obtains $\Delta\varepsilon_{\text{snow-oi}} = 0$. The bracketed [...] in Eq. (A1) is then equivalent with the permittivity of polycrystalline ice, $\varepsilon_{\text{ice}} = 1/3 \cdot \varepsilon_{\parallel} + 2/3 \cdot \varepsilon_{\perp}$, and it follows that $\varepsilon_{\text{snow-oi}} = f(\rho)$. This one-third/two-third weighting for polycrystalline ice has also been mentioned by Fujita et al. (1993, 2000) and was experimentally observed by Matsuoka et al. (1996).

For seasonal snow, some evidence has been found that the c -axis is preferentially vertically aligned for fresh snow (single-maximum) whereas for old snow the c -axis seems to be slightly oriented in a girdle in the horizontal plane (Riche et al., 2013). The strongest (single-maximum) anisotropy observed by Riche et al. (2013), parametrized by the eigenvalues $\lambda_z = 0.53, \lambda_x = 0.22$ result with Eq. (A2) in a maximum dielectric anisotropy of $\Delta\varepsilon_{\text{snow}} \approx 0.2 \cdot (0.22 - 0.53) \cdot 0.035 = -0.002$ for the example of $f(\rho) \approx f_{\text{vol}} = 0.2$ as common in seasonal snow. According to Fig.1(right) this must be compared to a structural anisotropy of $A = -0.02$ which is small

compared to the structural anisotropies observed in this paper.

Appendix B: Effective permittivity from weighted average of Maxwell–Garnett equations

5 The Maxwell-Garnett formulas (e.g. Eq. 3.27 in Sihvola, 2000) describe the effective permittivity of (elliptical) inclusion with a permittivity ε_i (e.g. ice) embedded in a host medium of permittivity ε_e (e.g. air) and can therefore be applied to calculate the anisotropic permittivity tensor for media with a structural anisotropy. The Maxwell-Garnett formulas can also be used in a "inverted" form, where the permittivities of inclusions and the host medium are swapped (e.g. air inclusions in ice). Both cases, the Maxwell-Garnett formula and the "inverse" Maxwell-Garnett formula, Eqs. (6a) and (6b), are equivalent with the lower and upper Hashin-Shtrikman-bounds for approximations of the exact description of dielectric mixtures (Hashin and Shtrikman, 1962; Sihvola, 2002).

For the isotropic case, the Maxwell-Garnett formulas can be compared with measurements of Mätzler (1996) and it has been found that the measurements lie well within the lower and upper Hashin-Shtrikman bound (Sihvola, 2002). However, the measurements are significantly larger than the lower Hashin-Shtrikman bound which is equivalent with the Maxwell-Garnett mixing formula for ice inclusion in air. Therefore, a combination of the Maxwell-Garnett formula and its inverse form is required for a better agreement with the measured data.

The measured data of Mätzler (1996) have been used in MEMLS-3 (Wiesmann and Mätzler, 1999, Eqs. 45/46) to fit an empirical formula to describe the permittivity of dry snow. The empirical formula is given in terms of snow density ρ and reads

$$\varepsilon_{\text{MEM-3}} = \begin{cases} 1 + 1.5995\rho + 1.861\rho^3 & \rho < 0.4 \text{ g/cm}^3 \\ \left[(1 - \nu)\varepsilon_h + \nu\varepsilon_{\text{ice}}^{1/3} \right]^3 & \rho > 0.4 \text{ g/cm}^3 \end{cases} \quad (\text{B1})$$

with the coefficients $\varepsilon_h = 1.005, \varepsilon_{\text{ice}} = 3.17$ (for -10°C) and $\nu = \rho/\rho_{\text{ice}}$. Note, that the original coefficients of $\varepsilon_h = 1.0, \varepsilon_{\text{ice}} = 3.215$ as given by Wiesmann and Mätzler (1999, Eqs. 45/46) have been adapted in agreement with C. Mätzler to produce correct results for pure ice ($\nu = 1$). We also note here that the exponent of $1/3$ is missing for the factor ε_s (here ε_{ice}) in Wiesmann and Mätzler (1999, Eq. 46).

Fig. B1(a) shows the lower and upper Hashin-Shtrikman-bounds (MG and MG,inv as dashed and dotted line) and the result of Eq. (B1) (solid line). The dots on top of the solid line indicate the weighted average of both bounds as given in Eq.(3). The weighted average is motivated by the fact that for low snow densities snow can better be described by ice particles embedded in a background matrix of air,

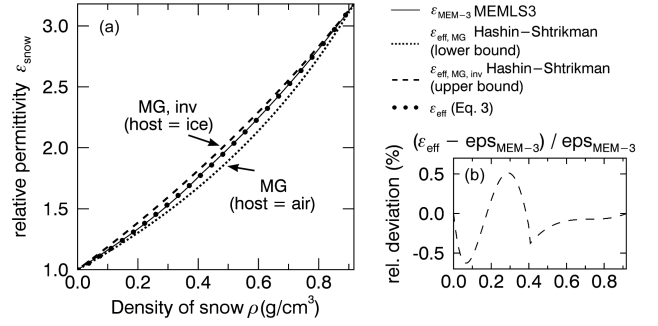


Figure B1. (a) Relative permittivity of snow $\varepsilon_{\text{MEM-3}}$ according to equation (B1) in comparison to the upper and lower Hashin-Shtrikman bound. (b) relative deviation of ε_{eff} (the weighted average of the Maxwell-Garnett formulas as given in Eq.(3)) with respect to the permittivity of dry snow as given by $\varepsilon_{\text{MEM-3}}$ in Eq. (B1).

whereas for high snow densities, snow can better be described by air inclusions in ice. Fig. B1(b) shows the relative deviation (in percent) between the permittivity used in MEMLS-3 (Eq. B1) and the weighted average of the Maxwell-Garnett formula and its "inverse" form. The relative deviation $(\varepsilon_{\text{eff}} - \varepsilon_{\text{MEM-3}})/\varepsilon_{\text{MEM-3}}$ is less than $\pm 0.7\%$ and justifies phenomenological composition of the averaged formula.

Appendix C: Re-derivation of Maxwell–Garnett equations via correlation functions

In Rechtsman and Torquato (2008) an exact series expansion of the dielectric permittivity of arbitrary anisotropic two-phase materials was derived and related to the n -point correlation functions of the material. If the series is truncated at $n = 2$, the final result (Rechtsman and Torquato, 2008, Eq. 16) can be solved for the diagonal components, $\varepsilon_{\text{eff},i}, i = x, y, z$, of the effective permittivity tensor which can be written in the form

$$\varepsilon_{\text{eff},i} = \varepsilon_q + \varepsilon_q \phi_p \frac{(\varepsilon_p - \varepsilon_q)}{\varepsilon_q + (1 - \phi_p) \left[\frac{1}{3} - \frac{U_i}{3\phi_p \phi_q} \right] (\varepsilon_p - \varepsilon_q)} \quad (\text{C1})$$

The permittivities and volume fractions of the two phases which compose the microstructure are denoted by $\varepsilon_p, \varepsilon_q$ and ϕ_p, ϕ_q , respectively. The quantities U_i in Eq. (C1) are related to integrals over the two-point correlation function $C(\mathbf{r})$ as defined in Löwe et al. (2013, Eq. 1). In the lowest order of frequency f , contributions from scattering in the effective permittivity can be neglected (cf. Rechtsman and Torquato, 2008, Eqs. C3, C4). Then the U_i have vanishing imaginary part and are given by

$$U_x = U_y = \frac{3}{4\pi} \int_{\mathbb{R}^3} d^3r \frac{1}{r^3} \left(-1 + \frac{3}{2} \sin^2 \theta \right) C(\mathbf{r}) \quad (\text{C2})$$

$$U_z = \frac{3}{4\pi} \int_{\mathbb{R}^3} d^3r \frac{1}{r^3} (-1 + 3 \cos^2 \theta) C(\mathbf{r}) \quad (\text{C3})$$

Here $r = |\mathbf{r}|$ is the magnitude of \mathbf{r} and θ denotes the angle between the vertical z axis and \mathbf{r} .

If the microstructure is (statistically) transversely isotropic, it is reasonable to assume a “spheroidal symmetry” of the correlation function, viz $C(\mathbf{r}) = C(r/\sigma(\theta))$ with $\sigma(\theta) = 2a_x[1 - (1 - a_x^2/a_z^2)\cos^2\theta]^{1/2}$ as used in Löwe et al. (2013). Under this assumption, the singular integrals in C2 can be calculated as shown in Torquato and Lado (1991). The results can be inserted into the square brackets in C1 leading to

$$\left[\frac{1}{3} - \frac{U_x}{3\phi_p\phi_q} \right] = Q \quad (\text{C4})$$

$$\left[\frac{1}{3} - \frac{U_z}{3\phi_p\phi_q} \right] = 1 - 2Q \quad (\text{C5})$$

where the anisotropy parameter Q is defined in Löwe et al. (2013, Eq. 4) or Torquato (2002, Eqs. 17.30/17.31). Using the definition of depolarization factors from Torquato (2002, Eq. 17.25), noting their relation to Q from Torquato (2002, Eq. 17.29) on one hand, and their equivalence to the definition of N_i from Eq. (8) and from the last paragraph of Sect. 2.2 on the other hand we end up with

$$\varepsilon_{\text{eff},i} = \varepsilon_q + \varepsilon_q \phi_p \frac{(\varepsilon_p - \varepsilon_q)}{\varepsilon_q + (1 - \phi_p)N_i(\varepsilon_p - \varepsilon_q)}. \quad (\text{C6})$$

We note here that Torquato (2002, Eq. 17.25) contains a typo. Specifying p to be the ice phase and q to be the air phase in Eq. (C6), gives $\varepsilon_q = \varepsilon_{\text{air}}$, $\varepsilon_p = \varepsilon_{\text{ice}}$, $\phi_p = f_{\text{vol}}$ in the notation from Sect. 2.2, and thus Eq. (C6) coincides with the Maxwell–Garnett result Eq. (6a).

Appendix D: CPD calibration of the SnowScat data

The measured radar signal was calibrated by an internal calibration loop of the SSI to compensate system drifts. However, some polarization dependent signal delay still originated from the connectors of the antenna feeding cables and from the antennas themselves due to the polarization-dependent beam-pattern. In order to calibrate external offsets and drifts, the CPD was calibrated with two metallic targets.

The primary calibration target was a metallic sphere with a diameter of 25 cm mounted on a wooden pole for the duration of the experiment. The sphere can be located in Fig. 5 next to the SSI. A secondary target, a metallic plate was located behind trees close to sector 2. A third calibration target, a dihedral reflector, was installed during the setup phase of the experiment. The correct pointing direction to locate the sphere was determined with a precision of $\pm 0.5^\circ$ by 2-D-scans in elevation and azimuth. The 2-D-scans showed that a possible systematic error of the CPD, caused by imprecise alignment, can be estimated to be below $\pm 10^\circ$.

The theoretical CPD measured from a sphere (or plate) is expected to be zero due to the target symmetry. The sphere was measured every four hours and was used as a reference during the whole duration of the experiment. The plate was installed from October 2011 until June 2013 and was used to validate the calibration done with the sphere. The CPD measured for a dihedral reflector should be 180° . The dihedral reflector was measured once, on 9 December 2009, to verify the processing sequence of the SnowScat raw data.

The CPD determined for the sphere, CPD_{REF} , was used as a reference and was subtracted from the uncalibrated CPD measurements of snow, $\text{CPD}_{\text{uncal}}$, to obtain calibrated results:

$$\text{CPD}_{\text{cal}}(f) = \text{CPD}_{\text{uncal}}(f) - \text{CPD}_{\text{REF}}(f). \quad (\text{D1})$$

Phase unwrapping was performed for the uncalibrated CPD and the reference CPD if necessary.

To reduce the noise of the reference measurements as much as possible, the reference, CPD_{REF} , was determined as follows: Time series $\text{CPD}_{\text{REF}}(t)$ were obtained for 21 different frequencies in order to sample the entire frequency spectrum between 9.2 and 17.8 GHz of the instrument. The time series were smoothed with a median filter of 4 days which preserved phase jumps in the signal. After temporal filtering, a frequency-dependent 4th order polynomial was fitted over the measured frequency spectrum of each acquisition to provide some noise reduction in the frequency domain.

The reference data are shown for all four seasons in Fig. D1. The solid black line shows the (frequency-dependent) reference, CPD_{REF} , for $f = 13.5$ GHz. Individual measurements of the sphere as well as measurements of the metallic plate are shown as dark and light gray solid dots below the black line.

In the third season, between 18 November 2011 and 20 January 2012, the pointing direction (elevation angle) to the sphere was misaligned by 2° . Therefore, the reference CPD was corrected by a frequency dependent offset to keep the CPD continuous at the start and the end of the period of misalignment.

The deviation of the raw-data of the sphere from the reference, $\Delta\text{CPD} = \text{CPD}(f) - \text{CPD}_{\text{REF}}(f)$, is shown in the lower panels for each season as scattered dots for each of the 21 analyzed frequencies. The root-mean-square-error, RMSE, was below 4° for the full frequency spectrum and is given for each seasons next to the graph. The error of the reference, $\text{CPD}_{\text{REF}}(f)$, which includes systematic and statistic errors, is estimated to be below 15° .

Appendix E: Author contributions

S. Leinss developed the birefringence model of snow, processed and calibrated the radar data, and wrote major parts of the manuscript. H. Löwe established the link between the Maxwell–Garnett theory and the statistical description of the

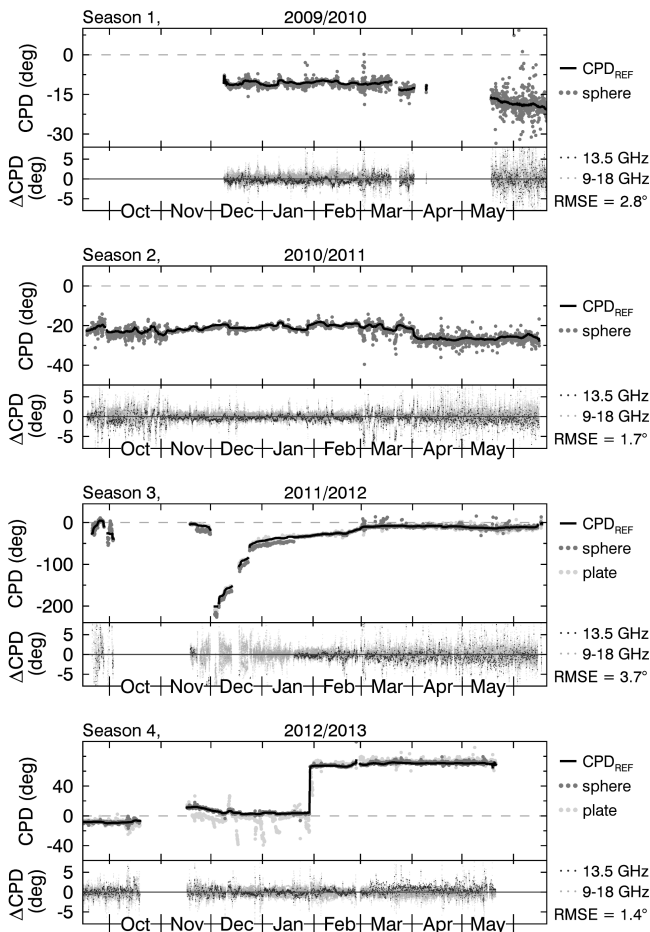


Figure D1. The CPD was calibrated using the sphere as a reference target. The upper panels show the reference, $CPD_{REF}(f)$, for $f = 13.5$ GHz (solid line) together with individual CPD measurements for the sphere and the plate (light and dark gray dots). The CPD of the metallic plate agrees within the standard deviation with measurements of the sphere and with CPD_{REF} . Deviations were found for season 3 due to a misalignment of the SSI to the sphere, and for November 2012, possibly due to snow cover of the metallic plate. The lower panels show the deviation, $\Delta CPD = CPD_{meas.}(f) - CPD_{REF}(f)$, for individual measurements at all measured frequencies $f = 10-17$ GHz. The deviation at a frequency of 13.5 GHz is shown as black dots. The standard deviation (RMSE) of ΔCPD for the whole frequency spectrum is given below the legend of the lower panel.

snow microstructure based on n -point correlation functions (Appendix C) and did the statistical analysis of the μ CT data. M. Proksch acquired the snow samples and calculated the anisotropy from the μ CT data. J. Lemmetyinen provided the snow measurements and meteorological data for the test site and added extremely valuable "in-situ" knowledge. A. Wiesmann added valuable knowledge about the processing and datasets of the SnowScat instrument. I. Hajnsek carefully proofread the paper and provided the funding for this paper.

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