Performance and experimental evidence of GPR in density estimates of snowpack

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ABSTRACT Ground probing radar (GPR) and microwave devices can be successfully used to analyse the structure and the density of the snowpack. The performances of standard radar systems for snowpack characterisation are analysed here; the main aim was to assess the reliability of the method for estimating snow density, snowpack thickness and the depth resolution in terms of capability of detecting thin layers. The main relationships between the electrical permittivity and the density of dry-snow are discussed; these relationships are applied to estimate the vertical density profiles inferred from GPR surveys performed at several test sites in the Italian Alps. The data are calibrated and compared with the results obtained from direct density and layer thickness measurements.

1. Foreword

The estimation of snow density and the water equivalent is a challenging task in hill-slope investigations. Snow water equivalent determines the amount of snow-melt discharge which is used in runoff modelling, and hence, to design drainage systems with on economic impact, if the snow-melt-water is to be used for power generation. As melted snow water carries pollution from snow dumps, snow density and water equivalent estimation are also related to the contaminant transport modelling. Finally, snow density has a direct effect on the mechanical strength of the snowpack and is, therefore, a parameter that should be considered in forecasting avalanches. The measurements of density are also relevant to validate the data obtained using remote sensing systems. Ground-based surveys, using portable ground-based equipment (TDR – Time domain reflectometry or GPR – Ground Probing Radar), can be useful to validate the information of remote sensing (InSAR) imaging of large areas.

The mechanical properties of the snowpack are usually inferred by using conventional approaches to estimate the density, the cohesion and the free (liquid) water content; for instance static penetrometers are adopted to estimate the consistency through the measurement of the depth to which a standard needle penetrates into the snowpack.

The measurement of vertical density profiles assumes great importance as spatial changes and time-varying of density along vertical profiles must be considered to estimate the avalanche risk (Conway and Abrahamson, 1984). Furthermore, different degrees of snow metamorphism can be recognised through density estimation.

Combined high (using TDR) and low-frequency permittivity measurements are usually performed for continuous snow wetness and snow density determination (Stacheder *et al*., 2005);

new electromagnetic sensors for large-scale snow-cover monitoring are continuously being developed (Stähli *et al*., 2004). These systems allow one to estimate the real and imaginary part of the electrical permittivity at different frequencies; the complex permittivity is then related to density values of the dry or wet snow using empirical rules or mixing models. An application of the model requires accurate calibration to take into account the effect of thermo-dynamical processes. However, the application of new-generation sensors is spatially-limited to restricted and well-controlled areas (Stähli *et al*., 2004). On the other hand, the georadar investigation can cover large areas and work in unfavourable logistical conditions.

This paper gives a short description of the relationships between the electrical permittivity and snow density; the theoretical background is followed by the interpretation, in terms of vertical density profiles, of the results of ground probing radar surveys conducted at several sites, characterised by different snow conditions, in the Italian Alps. The results of the georadar interpretation were calibrated and compared with the density estimate using a conventional approach.

The reliability that high frequency radar devices have, particularly stepped-frequency systems, to relate the mechanical properties (density and strength) of thin layers to the georadar response has been proven by Marshall *et al*. (2007); on the other hand, the application to large areas is not yet well documented. Impulse radar systems offer a robust low cost tool for mapping wide areas to explore the behaviour of the snow cover with relevant thickness (e.g. more than 3 meters). The experimental results discussed here show that impulse radar systems are capable of pointing out and assessing the lateral continuity of interesting features in the snowpack, such as ice crusts, which are relevant in avalanche risk estimation.

2. Snow density and electromagnetic models

Density-forming processes can be classified as primary, that develop at the snow surface either during deposition or through subsequent reworking due to wind; and, secondary, that evolve on a site subsequent to deposition and are due to thermo-mechanical processes.

Primary densification is a function of atmospheric conditions, precipitation rate, size and type of snow crystals, and the packing and disintegration of crystals by the wind. Secondary densification results from a volumetric creep due to normal and shear stresses, and from the linked processes of heat and mass transport.

Snow density depends on the grain typology, snow temperature and winds; the density (ρ_m) can be computed according to the contribution of each component:

$$\rho_m = \left[\rho_{ice} \cdot (1-\phi) + \rho_{water} \cdot \theta\right] \tag{1}$$

where ϕ is the snow porosity, θ is the free-water content, ρ_{ice} and ρ_{water} are the density of ice and water, respectively. The water content depends on the temperature and transformations of the snow pack. Fresh snow has a very low free water content (less than 3%); in spring, during snow melting, the water content can reach values of more than 15%. An increase in the mass, for instance due to fresh snow precipitation, leads to a compaction of the layers with a reduction in the air volume or the melting of old snow and transformation into ice or liquid water. The air temperature during snowfalls influences the density of fresh snow; density ranges from 20 kg/m^3 for temperatures of –20 Celsius up to 250 kg/m^3 for temperatures close to zero. The fresh snow density is related to other factors such as the shape of the snow crystals and the temperature gap between the crystals and the ground. Finally, the thickness and the compaction of old snow can modify the thermal circulation between subsoil and water and could influence the metamorphism of the snowpack.

The electrical conductivity of snow depends on the porosity which is inversely proportional to the bulk density, the morphology of the snow crystals, the metamorphism, the water content and temperature. For density ranges between 240 kg/m^3 (fresh snow) and 640 kg/m^3 (granular snow), Kopp (1962) showed that the electrical conductivity ranges between 10^{-9} to 10^{-7} S/m at a temperature of –10º Celsius. An increase in the conductivity can be observed at a temperature close to 0º Celsius where values of 10^{-5} S/m have been measured.

Electrical permittivity indicates the polarizability of a material: as the frequency increases, the polarisation response may lag behind the varying field; therefore, the general form of permittivity involves an in-phase (ε') and out of phase (ε'') component:

$$\varepsilon = \varepsilon' - i\varepsilon''$$

where ε'' is also called the loss factor. The loss factor of snow depends on the percentage of the free liquid water content and can vary by several orders of magnitude for snows with different amounts of liquid water.

The electrical permittivity can be estimated according to the assumption that snow is a mixture of air, water and ice; the permittivity of dry compact snow (with a water content below 3%) is very similar to the dielectric value of ice [ε_{ice} =3.2: Evans (1965), Mätzler (1996)].

Dry-snow can be considered as a non-conducting medium; the electromagnetic wave does not suffer from intrinsic attenuation as it propagates through the snowpack, therefore, it can be assimilated to a lossless medium; in such a case the complex permittvity ε_c is equal to the real permittivity ε.

The wavenumber k is defined by:

$$k = \omega\sqrt{\mu \cdot \varepsilon}$$

and the phase-wave velocity is:

$$v = \frac{\omega}{k} = \frac{1}{\sqrt{\mu \cdot \varepsilon}} \qquad (2)$$

where μ is the magnetic permeability of the medium ($4\pi 10^{-7}$ H/m in the S.I. unit) and ε is the dielectric constant of vacuum (8.852 10^{-12} F/m).

The intrinsic impedance is:

$$\eta = \sqrt{\frac{\mu}{\varepsilon}} \qquad [\text{ohm}] .$$

The phase velocity and the wavelength are related by:

$$\lambda = \frac{2 \cdot \pi}{k} = \frac{v}{f} \qquad [\text{ m }]$$

the wavelength (λ) of the travelling wave is proportional to the phase velocity and inversely to the frequency (f).

For a normal plane wave incidence at the boundary between two snow layers or between the snowpack and the bedrock, the following reflection (Γ) and transmission coefficient (τ) are considered:

$$\Gamma = \frac{\eta_2 - \eta_1}{\eta_2 + \eta_1}$$
$$\tau = \frac{2\eta_2}{\eta_2 + \eta_1} \tag{3}$$

and for non-magnetic materials (such as snow):

$$\Gamma = \frac{\sqrt{\varepsilon_1} - \sqrt{\varepsilon_2}}{\sqrt{\varepsilon_1} + \sqrt{\varepsilon_2}}. \tag{4}$$

The performance of the radar survey can be estimated a priori, considering the response of a single, thin layer as a rough planar target radar response. In such a context, the target is defined as a rough planar target when the ratio between the surface roughness and the wavelength is greater than $1/2\pi$. The wavelength in the frequency range between 500 MHz and 1000 MHz is 0.2 m and 0.1 m, respectively. Therefore, it is appropriate to consider the thin snow layer as a rough planar target s especially when high frequencies (> 900 MHz) are adopted.

The radar cross-section of a planar rough targets can be assumed as the First Fresnel, defined by $\Gamma^2\pi\lambda D/2$, where D is the distance object-antenna distance (in the single reflection zero-offset mode) and Γ^2 is the plane wave power reflection coefficient. The maximum penetration depth can be determined by taking into account the performances, the sensibility of the radar system and the antenna gains. From a theoretical point of view, a good compromise between penetration depth (more than 3 meters) and vertical resolution could be obtained at the main frequency of 900 - 1000 MHz.

The vertical resolution is the minimum vertical distance between two interfaces (e.g. snow layers with different density) that can be identified in the radar images as two well-distinguished events. This depends on the radar performances and the wavelength of the signal; therefore, as a secondary parameter, it also depends on the density of the snow pack. For instance, a high granular-density snow (> 600 kg/m^3) is characterised by a wavelength of 0.2 m (at 1 GHz) and a theoretical vertical resolution of 0.05 m (1/4 of the wavelength).

When a thin layer of a medium 2 is embedded into a medium 1 and the thickness t of the layer 2 is comparable to the wavelength in the first medium, a more appropriate expression for the reflection coefficient is (Annan *et al*., 1988):

$$\Gamma = \frac{\Gamma_{12} \cdot \left(1 - e^{i\beta}\right)}{1 - \Gamma_{12}^{2} e^{i\beta}} \tag{5}$$

where $\beta = \dfrac{4 \cdot \pi \cdot t}{\lambda_2}$ where λ_2 is the wavelength in medium 2.

In order to estimate the theoretical depth resolution (ΔR), the bandwidth (B) and the shortest wavelength λ_s, corresponding to the upper frequency bound f_u of the received signal bandwidth, must be considered:

$$\Delta R = \frac{v}{2B} = \frac{f_u \cdot \lambda_s}{2B}. \tag{6}$$

As far as the correlation between the electromagnetic parameters and the snow density are concerned, different approaches can be considered; a simple standard model is the Robin equation (Robin, 1975) which is an empirical relationship between density and electrical permittivity:

$$\varepsilon_{r,ice} = \left(1 + 0.845 \cdot \rho\right)^2. \tag{7}$$

A mixing model could be adopted to describe the macroscopical behaviour of the electrical permittivity of a mixture of ice, air and water (Birchak *et al*., 1974):

$$\varepsilon = \left[\varepsilon_{ice}^{0.5} \cdot (1 - \phi) + \varepsilon_{air}^{0.5} \cdot (\phi - \theta) + \varepsilon_{water}^{0.5} \cdot \theta\right]^2 \tag{8}$$

where the permittivity of the mixture (ε) is the weighted average (in volume) of the combination of permittivity of the snow crystals (ε_{ice}), water (ε_{water}) and air. A simple model used to estimate the dielectric behaviour of dry snow with respect to the density was proposed by Looyenga (1965):

$$\varepsilon_{r,snow}^{1/3} - 1 = x \cdot \left(\varepsilon_{r,ice}^{1/3} - 1\right). \tag{9}$$

In this study the following relative values of electrical permittivity are considered: ε_{ice}=3.2, ε_{water}=78 (at 0º Celsius) and ε_{air}=1; the density value of the snow crystals is equivalent to 920 kg/m^3 (Glen and Paren, 1975).

Where $x = \frac{\rho_{mixture}}{\rho_{ice}}$ is the ratio between the density of the snow mixture and the ice (920 kg/m^3).

The theoretical relations between the density and electromagnetic parameters of the snowpack are estimated and plotted in Fig. 1, according to the aforementioned formulas. Table 1 shows some theoretical values of the electromagnetic parameters for different types of snow.

Fig. 2 plots the trend of the absolute value of the reflection coefficient, computed according to Eq. (5), for a thin layer of an ice crust (ε=3) embedded in a softer snow, with a permittivity value of ε=2. The general trend of the reflection versus frequency is affected by the thickness of the layer; in the selected frequency band, a linear approximation is obtained for very thin layers (t= 5–10 mm). As the layer thickness increases, the reflection coefficient assumes sinusoidal behaviour with a maximum of the reflection coefficient at different frequencies: the peak for a 50 mm thin layer is centred at the frequency of 1 GHz. The reflection coefficient of the 100 mm-thick layer is almost null at the frequency of 1 GHz; at some particular frequencies (related to 1/4 - 1/2... of the wavelength) the reflection of the layer with a thickness of 100 mm could be negligible. As far as the amplitude of the reflection coefficient is concerned, the values increase from 0.025 for a 5 mm-thin layer up to 0.05 for a layer of 10 mm and to 0.2 for a 50 mm-thin layer (frequency 1 GHz).

3. Methods

Radar surveys have been conducted for snow thickness detection for over 30 years; in northern Europe, radar track-mounted systems have been employed for snow thickness evaluation and to estimate the equivalent water content of snowpacks (Ulriksen, 1982) for hydrology purposes and the

Table 1 - Theoretical values of electromagnetic properties for different snow densities (1 GHz).

Snow	Density range [kg/m^3]	Theoretical Wave velocity* [m/s] ·10^8	Electrical Permittivity	Wavelength [m]	Impedance [ohm]
Freshsnow	100 300	2.78 2.42	1.16 1.54	0.28 0.24	349.4 303.7
Granular snow	400 600	2.27 2.00	1.75 2.24	0.27 0.20	284.5 251.6
Ice crust	700 900	1.89 1.69	2.52 3.13	0.19 0.17	237.4 212.8

* according to Eq. (9)

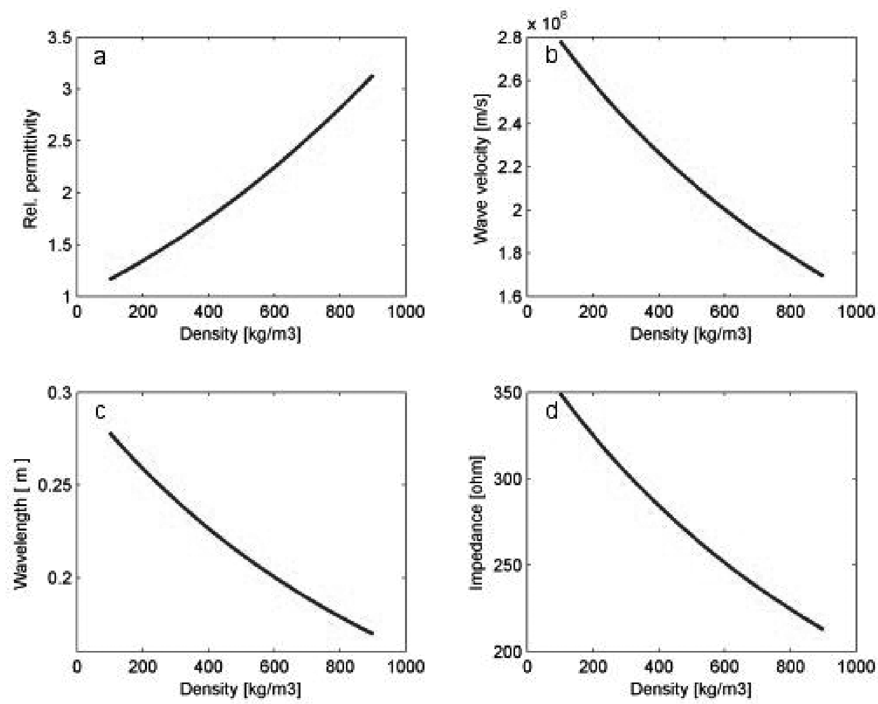

Fig. 1 - Theoretical trend of the electrical permittivity of snow (a), wave velocity (b), wavelength (c) electrical impedance (d), according to the Looyenga (1965) model (see text).

recharging of shallow aquifers. Experiments conducted to detect human bodies buried by avalanches were carried out in the 1980s and 1990s, with ambiguous results because of the practical limitations due to the slowness of the radar survey over large and rugged areas. Radar methodology in snowpack analysis should provide information at different scales according to the adopted approach and the main frequency of the antenna (Annan *et al*., 1994; Harper and Bradford, 2003).

A single reflection mode with zero offset or constant offset is the conventional approach for snowpack layering and thickness estimation; it is reasonable to assume that only an estimation of the average (on the snowpack) density and water equivalent content can be carried out. A vertical cross-section of the snow cover is obtained by moving the transmitter/receiver device along a transect. The radar image displays the amplitude of the reflected wave versus two-way travel-times and the antennas position along the surface. The two-way traveltimes can be converted into depths if the wave velocity can be estimated (for instance according to snow thickness measurements using a micro-penetrometer device). However, only in a medium with uniform density, can the two-way traveltime be converted into a depth by assigning one average velocity without introducing significant errors in the depth estimation. On the other hand, a uniform density and constant water content is seldom detected in the field; in layered snowpack where thermo-mechanical processes are responsible for intense snow transformations, the hypothesis of constant density can, in particular, cause relevant inaccuracies.

The multifold radar (CMP or CDP) survey provides a better vertical resolution and more

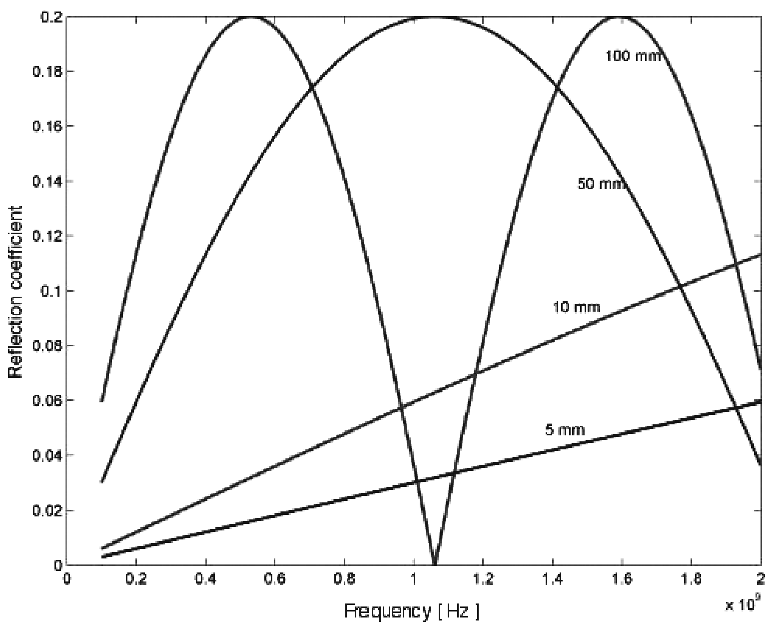

Fig. 2 - Amplitude of the reflection coefficient in the frequency domain due to a thin ice-crust layer (electrical permittivity ε=3) embedded in a softer snow (ε=2); the thickness of the thin layer is indicated on the plot.

detailed investigation than the single reflection approach, but it is much more costly and time expensive. Therefore, the multifold investigation is rarely used on hill-slopes to map a snow cover over large areas.

The spatial resolution is mainly associated to the radar wavelength and to the stratigraphic snowpack conditions; CMP or CDP data acquisition using a multifold approach can provide information on the density and water content through the analysis of the interval velocity between two reflection interfaces.

The vertical resolution of the multifold approach is related to the presence and separation between horizontal interfaces and their detectability; the wave velocity derived from this approach refers to an average value computed according to the thickness of each horizon. The method is more suitable to estimate the lateral variability of densities at different depths than to offer a high resolution in vertical discrimination. Using the WARR approach (Wide Angle Refraction and Reflection), it is possible, from a theoretical point of view, to determine both the ground wave velocity and the interval velocity due to the reflection at the snowpack interfaces.

The data quality must be assessed taking into account the experimental inaccuracies, such as the antenna positional error, the time picking uncertainty, the time shift, the disturbance of the radar signal due to snow surface roughness and subsurface features (snow impurities and rock debris) close to the antenna.

The radiation pattern and the Fresnel radius determine the cross-sectional area of the interfaces that intercept and reflect the radiated power; the first Fresnel radius is plotted (Fig. 3) versus the two-way traveltime for a reference wave velocity of 0.2 m/ns at a frequency of 1 GHz. It can be seen that at the traveltime of 10 ns, approximately at a depth of 1 m, the radius is equal

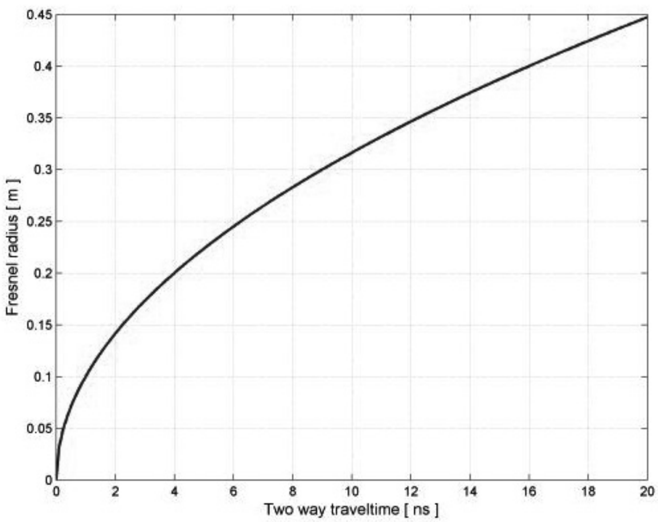

Fig. 3 - The first Fresnel radius versus the two-way traveltime for a 0.2 m/ns wave velocity at 1 GHz.

to 0.3 m; at a depth of 2 meters (twt=20 ns), the Fresnel radius increases to 0.45 m. These values should be kept in mind when the lateral resolution of the radar performance is dealt with or when the effects of the snowpack-ground morphology on the radar response have to be considered in estimating the snow cover thickness.

4. Georadar experiments

The experiments were conducted in the Sestriere sky area in the western Italian Alps in the Olympic Valley on the hill-slopes near Sestriere and Sauze Oulx; the surveys consisted of radar investigations and comparisons with the stratigraphic evidence obtained from the vertical profiles of the mechanical properties of the snow covers.

The conventional approach for monitoring the mechanical properties of a snowpack is based on the excavation of trenches or pits (2-3 meters or more in depth, according to the maximum thickness of the snow cover) where the free water content, the metamorphism of the snow crystals and the density are evaluated at different depths. Static penetration tests are usually performed to estimate snow compaction and cohesion.

Several tests were performed in different areas at mid-elevation (between 2000 and 2500 m a.s.l.) in the months of February-March; they, therefore, refer to the deposition and transformation of fresh snow during the winter. Snow falls are frequent in the area, starting from November; the snow reaches maximum values of 4-5 meters in the thickness pack at an elevation of 2500 m a.s.l.. The metamorphism of the snow is usually intense due to the climatic conditions of the area; the winds significantly modify the snow conditions and cause mechanical discontinuities of the snowpack. These climatic conditions determine different types of snow transformation during the winter and the formation of thin crusts of ice which, with particular slope conditions, can cause avalanches.

The GPR surveys were performed using different antennas in the frequency band ranges from 500 MHz and 1500 MHz; data were acquired both in single fold and multifold configurations. The results of surveys performed at three different sites (Vallonas, Broussailles, and Fraiteve) are discussed.

The data processing of the single fold reflection data involved:

- marker interpolation to normalize the horizontal trace distance at an interval of 0.125 m;
- time picking of the reflected signal at the snow-ground interface;
- estimates of the average wave velocity according to the data processing of the CDP data;
- depth conversion using the wave velocity detected from the CDP data.

Finally, the Hilbert-transform is used to calculate the instantaneous attribute (envelope), which gives a measure of the reflectivity strength; the envelope gives an overview of the energy distribution of the traces: it can facilitate the determination of signal, first arrivals of the snow-ground reflection and point out the presence of the weak reflection due to the snow layering.

A preliminary estimate of the depth resolution considers that the vertical resolution is related to the bandwidth of the B-signal received and the shortest wavelength of the signal bandwidth received; the spectral analysis performed on the signal part of the reflected waveform made it possible to estimate the average wavelength values of 0.24 m (500 MHz) and 0.13 m (900 MHz): the quarter-wavelength calculation yielded a theoretical vertical resolution of about 5-10 cm.

The Vallonas test site is located on a perfectly planar ground in a small isolated valley, where the snow accumulation is not affected by wind or human activities. Metamorphism is, therefore, mainly due to the temperature changes during the winter. The thickness of the snowpack was measured at different points along the profiles surveyed by the GPR at the main frequency of 900 MHz; an average thickness of 1.40 ± 0.05 m was estimated using a calibrated steel rod. The GPR single fold profiles made it possible to estimate the two-way traveltimes of the substratum reflection with an accuracy of 0.5 ns (Figs. 4 and 5). This leads to an average velocity value of 0.235 m/ns, which corresponds to an average snowpack density of 340 kg/m^3 , according to Eq. (9). The measurements of the free water content and density performed on the trench showed out a very low free water content; this condition permitted us to estimate the average density of the snowpack, starting from the traveltime radar measurement and applying Eq. (9). A weak reflection at a depth of about 0.7 m from the surface was observed. The presence of this reflection is enhanced by the envelope analysis: this corresponds to an increase, in the density, from a value of 340 kg/m^3 to 410 kg/m^3, as estimated from the measurements performed in a pit excavated in the snow.

The radar data acquired at the Broussailles site are plotted in Figs. 6 and 7; the images show two vertical sections of the snowpack acquired in single reflection modes: the raw data are at the top and a filtered and processed image is at the bottom. The air-snow interface provides an intense reflection signal in the raw data; the contrast of the electrical properties between the snowpack and the ground determines a strong reflection signal; weaker reflections, related to the snow layering, can be detected inside the snow pack, as pointed out by the envelope analysis.

The snow cover is characterised by three main macro-layers, that can be distinguished for different metamorphism and density values in the range between 290 kg/m^3 and 425 kg/m^3, according to the measurements performed with a dynamometer on samples collected at different depths in a snow pit. The snow density increases from the surface to the bottom of the snowpack,

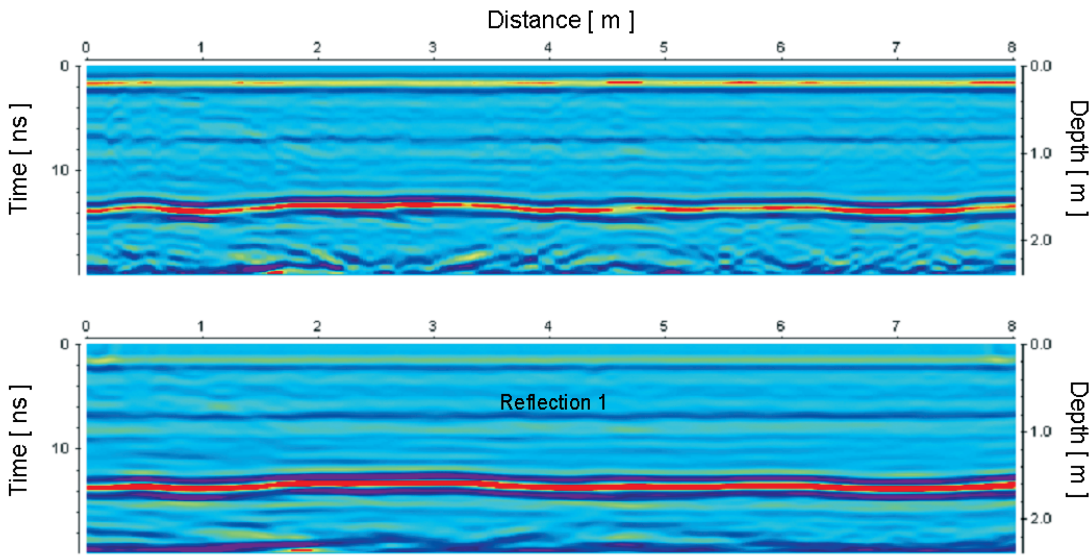

Fig. 4 - The Vallonas site - Profile 1: top) raw radar data collected using a central frequency of 900 MHz; bottom) filtered and migrated image; the main reflection at 13-14 ns refers to the snow-ground interface; reflection 1 is related to the presence of crust winds; snow layering is also clearly detected.

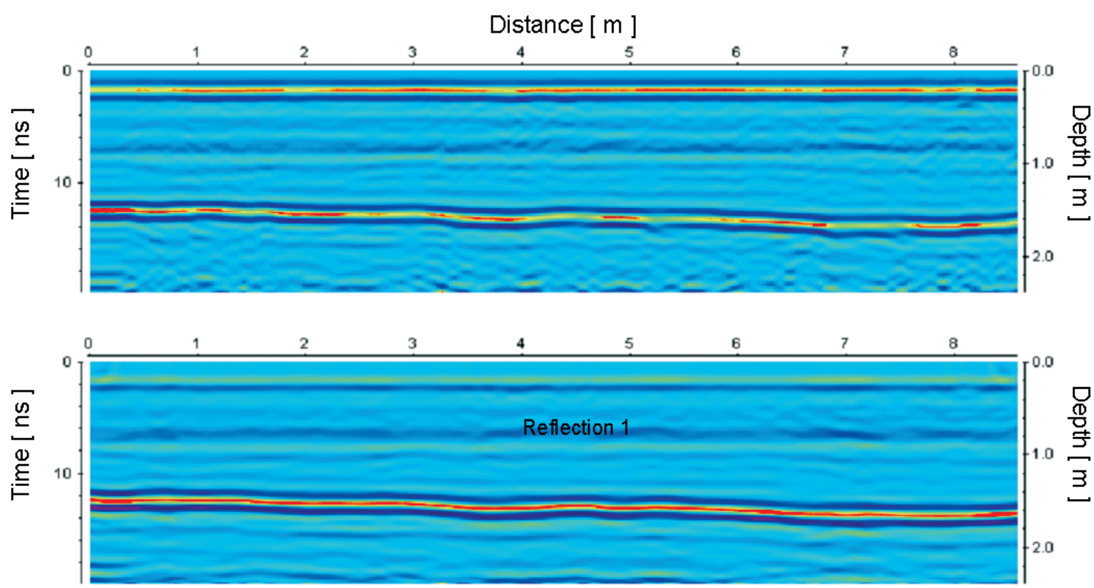

Fig. 5 - The Vallonas site - Profile 2: top) raw radar data collected using a central frequency of 900 MHz; bottom) filtered and migrated image; the main reflection at 13-14 ns refers to the snow-ground interface; the crust winds (reflection 1) appear less continuous than the image in the previous figure; snow layering below reflection 1 is also clearly detected.

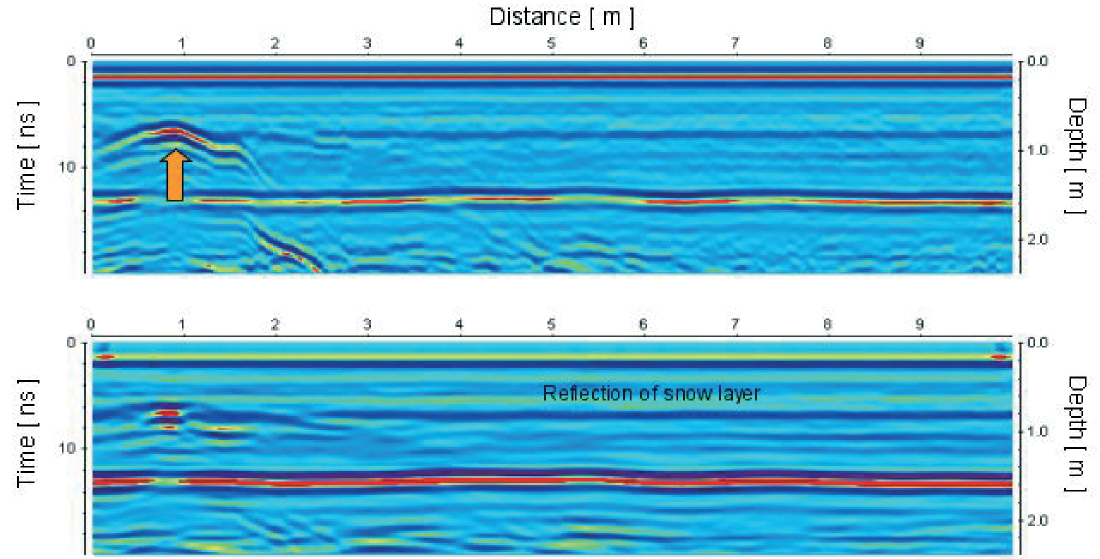

Fig. 6 - The Broussailles site - Profile 1: top) raw radar data collected using a central frequency of 900 MHz; bottom) filtered and migrated image; two skis at 0.65 meters in depth, buried on the snow from a lateral trench, determine the reflection on the left side of the radargrams (marker A).

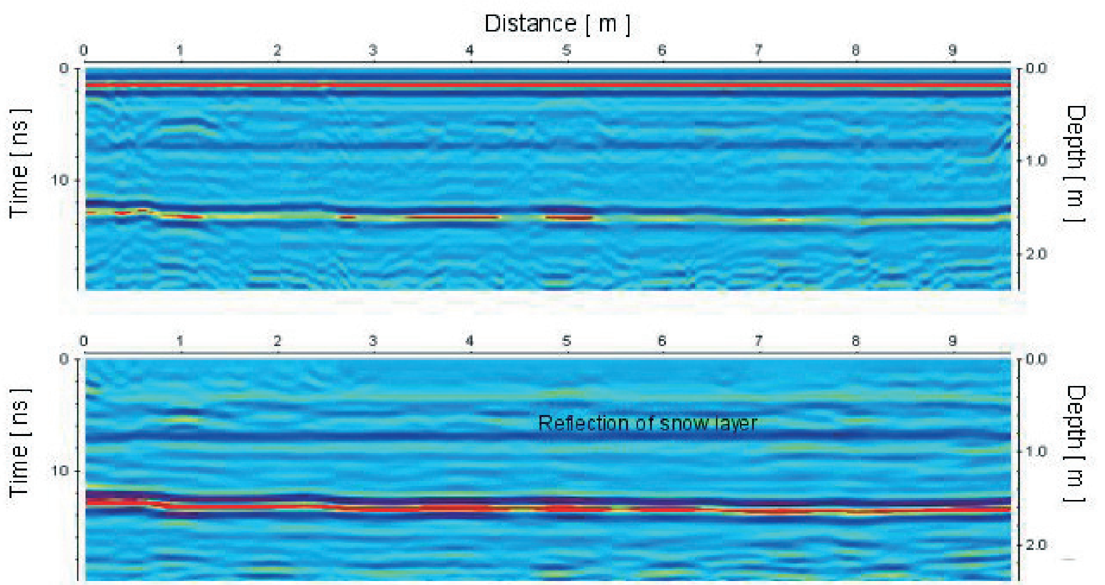

Fig. 7 - The site Broussailles - Profile 2: top) raw radar data collected using a central frequency of 900 MHz; bottom) filtered and migrated image.

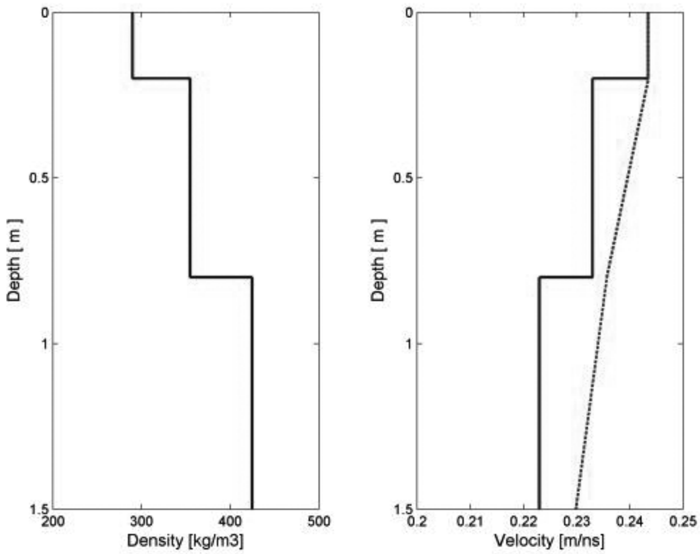

Fig. 8 - Left) density vertical profile of macro-layers as determined by the density evaluation within a snow pit; right) corresponding estimated values of interval velocity and root mean square velocity.

with a different compaction and metamorphic transformation.

The velocity analysis performed on the CDP radar data made it possible to estimate the electromagnetic wave velocity in the range between 0.23 m/ns and 0.24 m/ns, which corresponds to the electrical permittivity range between 1.5-1.7 (Figs. 8 and 9).

The uncertainty in the density evaluation can be assessed starting from the accuracy in the electromagnetic wave velocity performed on the CDP data. An average value of 0.235 ± 0.005 m/ns yields an estimate of the permittivity value of 1.63 ± 0.08: the equivalent density values are in the range of between 310 kg/m^3 to 380 kg/m^3, according to the Looyenga (1965) model. The discrepancy between the density inferred from the gravimetric measurements (with a dynamometer) and the ones estimated using the radar data can be put down to the inaccuracy in determining the average density value of the whole snowpack and the impossibility of estimating the density of each layer from the radar data; moreover, the sensitivity of the relationship between the density and electrical permittivity is rather high in the selected range and causes an increase in the uncertainty of the density estimation, as discussed in the next section.

The Fraiteve site is located at 2550 m a.s.l.; the area is subjected to strong winds that are responsible for a significant amount of snow transport and accumulation. The site usually presents a higher thickness of snow than the rest of the slope and formations of anomalous crust winds above low-consolidated snowpacks. The slope morphology and the local climatic factors are responsible for a high avalanche risk. The GPR survey had two main aims: to check the reliability of a fast estimate of the snowpack thickness on the top of the slope and to verify the presence and the lateral continuity of ice-crusts. The radar survey was performed at 500 MHz and 900 MHz in single fold (Figs. 10 and 11) and CDP acquisition for the wave velocity estimate.

The effectiveness of the radar survey was verified along two profiles by means of direct measurements of the snow thickness using a calibrated rod; a maximum thickness of the snow cover of about 2.5 meters was detected. Eq. (9) between the snow density and electromagnetic

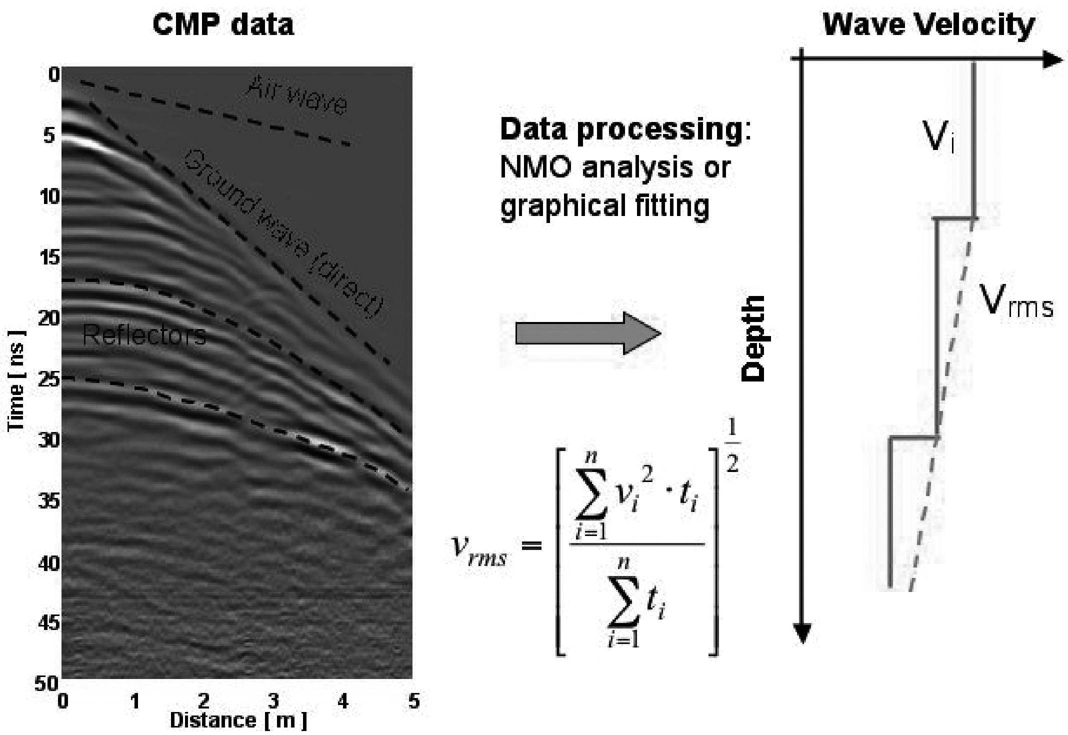

Fig. 9 - Example of CDP acquisition for interval velocity or root mean square velocity estimate of the snowpack.

wave velocity and electrical permittivity was applied to determine the average value of density of the snow cover to be estimated. A trench permitted the stratigraphic sequence of the snow cover to be estimated; the presence of ice crusts was detected at depths of 88-90 cm and 112-115 cm from the snow surface. These crusts separate low-consolidated snow layers characterised by low density, with a reduced free water content and slight metamorphic transformation.

5. Discussion and conclusion

The snowpack at the test sites was mainly characterised by dry-snow with a very low free water content; this makes it difficult to extend the results to more general snow conditions, where the melting process could cause an increase in the free water content. In such cases, the estimate of density values from velocity analysis could be ambiguous, because of the of lack of reliability in the density-permittivity model adopted here. However, with these conditions, the radar survey permitted a fast estimate of the overall snowpack cover and a qualitative analysis of the different phases of deposition (snowfalls or wind transport). They were clearly identified, above all, in the single-fold images acquired at the Fraiteve test site, where the older layers (close to ground) realistically reproduced the morphology of the substratum, as can be observed from the analysis of the reflection events related to the snow layering.

The experimental results pointed out:

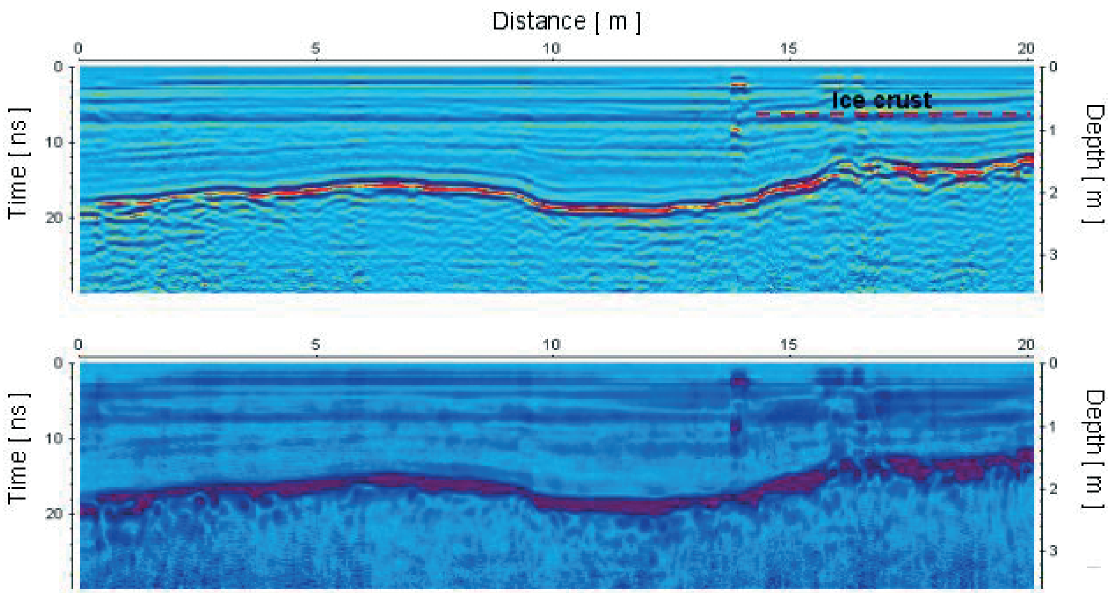

Fig. 10 - The Fraiteve site - Profile 1: top) radar image acquired at 900 MHz; bottom) image of the instantaneous amplitude envelope to enhance the weak reflection due to snow layering.

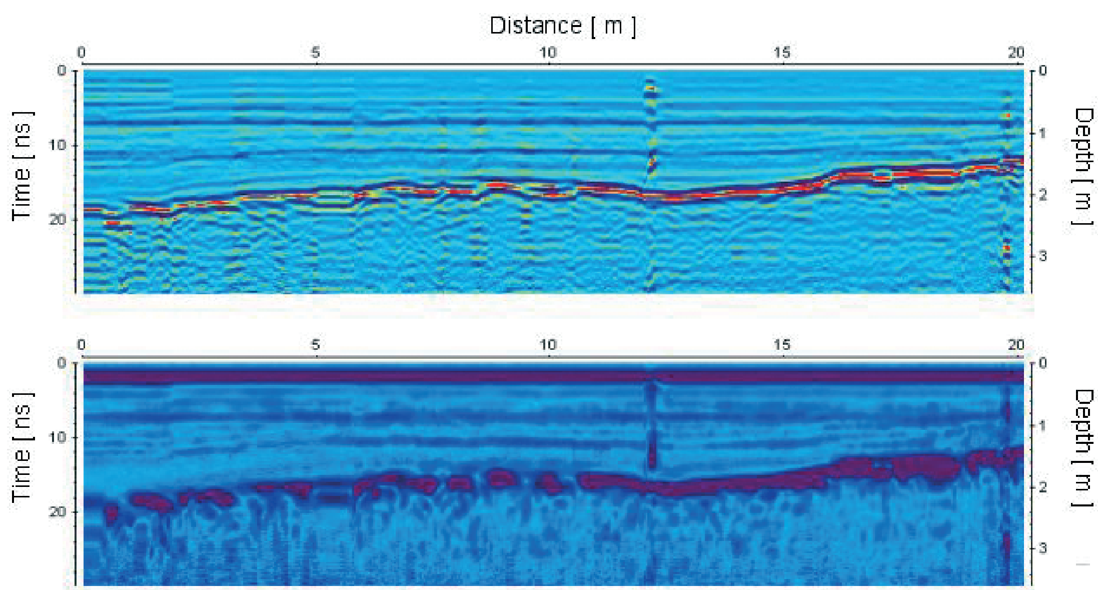

Fig. 11 - The Fraiteve site - Profile 2: top) radar image acquired at 900 MHz; bottom) image of the instantaneous amplitude envelope to enhance the weak reflections due to snow layering

- a good correspondence between the radargrams and the snowpack thickness; a good lateral resolution in determining irregular snow deposition can be achieved using the 900 - MHz antenna; in fresh snow or in snow with low water content, the penetration depth of the radar survey was excellent: with a high frequency antenna, a snowpack of up to 4 meters in depth can be analysed;
- unfortunately the high velocity of dry snow and, therefore, the high wavelength does not permit a satisfactory vertical resolution for accurate detection of different layers;
- the discrepancy between the density estimated on samples collected in the snow pits and the results of radar survey is mainly due to heterogeneity of the snowpack, inaccuracy in the gravimetric measures, local terrain irregularities and average density assumption for snowpack evaluation from radar data;
- the weak reflections within the snowpack are related to ice-crusts, whose presence were confirmed from the analysis of the stratigraphic profile evaluated in the snow pits;
- the radargrams showed the seasonal deposition and compaction of different snow layers; the first snowfalls and wind accumulations assume a very similar trend to the terrain morphology; later accumulations are characterized by the radar reflection with a similar trend to the snow-air interface, generally, when the snow coverage assumes a relevant thickness.

The estimate of the error propagation is performed considering the average wave velocity inferred from the CDP measurements; for instance, for an average wave velocity equal to 0.2 ± 0.01 m/ns, the interval of the electrical permittivity estimate is:

$$\varepsilon_r' = \frac{c^2}{(v \pm \Delta v)^2} = \frac{0.3^2}{(0.2 \pm 0.01)^2} = \begin{Bmatrix} 2.04 \\ 2.49 \end{Bmatrix}$$

for an average wave velocity equal to 0.24 ± 0.01 m/ns,

$$\varepsilon_r' = \frac{c^2}{(v \pm \Delta v)^2} = \frac{0.3^2}{(0.24 \pm 0.01)^2} = \begin{Bmatrix} 1.44 \\ 1.70 \end{Bmatrix}.$$

The uncertainty propagation in the density estimation is computed for the Looyenga (1965) model, as showed in Fig. 12; the plot shows the ranges of the density values, starting from the range of the estimate of the electrical permittivity. It can be seen that for the average velocity of 0.2± 0.01 m/ns the density values are in the 500 kg/m^3 to 700 kg/m^3 range, with an uncertainty of about 16%; at the average velocity of 0.24 ± 0.01 m/ns, the density can be estimated in the 220 kg/m^3 and 380 kg/m^3 range, with an uncertainty of about 26%.

The joint application of a radar survey (single fold mode) and the mechanical measurements of snowpack thickness makes it possible to estimate the average values of density with accuracy of 5-10% for snowpacks in the range of 2-3 meters. Favourable snow layering conditions, for instance, alternation of ice-crusts and low-consolidated snow layers, could detect the spatial

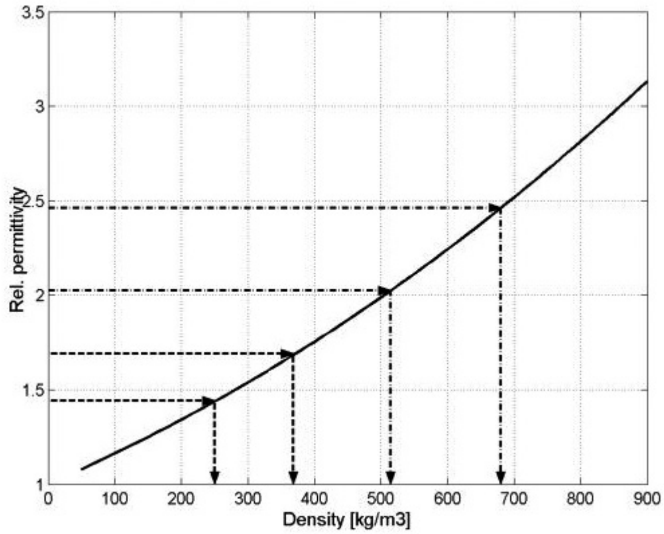

Fig. 12 - Sensitivity of the Looyenga (1965) model in the density estimate starting from the accuracy in the electrical permittivity estimate; the dashed lines refer to the interval of the density starting from a velocity value of 0.2 ± 0.01 m/ns; the dot-dashed lines refer to a velocity value of 0.24 ± 0.01 m/ns.

coherence of the ice layers.

Multifold data acquisition and velocity analysis could offer an improvement in determining the vertical density profile; a more accurate detection of snow layering and an estimate of the changes in the mechanical properties of the layers can be achieved using higher frequencies (> 2 GHz), that can be adopted in dry snow.

6. Final remarks

The radar survey in single-reflection mode, together with mechanical measurements of snowpack thickness, allows one to estimate the average density values of the snowpack with an accuracy of 5-10% for a thickness of the cover of up to 3 meters. The method could qualitatively permits the detection and spatial coherence of ice layers or, more in general, to assess the snow layering inside a snowpack and to detect the alternation of ice crusts and low consolidated snow layers. The velocity analysis using the CDP approach can be successfully adopted to quickly estimate the trend of a vertical density profile. The estimate of the average value of density can be obtained with acceptable accuracy; the problematic task of the micro-structure properties of the snow cannot be solved.

Several relationships can be adopted to convert the electromagnetic parameters to the density or the water equivalent snow parameters; mixing formulas can be considered for dry snows under low-density conditions; wet snow exhibits a relaxation spectrum in the radiowave-microwave range, and more realistic physical mixing models for small values of the free water content are necessary (e.g. Mätzler, 1987).

The low water content (below 3%) of the snow at the selected sites in the Italian Alps, permitted us to adopt simplified mixing formulas with satisfactory results. However, a much bigger effort should be performed in order to determine more robust and accurate relationships between electromagnetic parameters and the micro-structure of snow. For instance, the imaginary part of the dielectric permittivity and frequency dispersion could be related to the structure of wet snow.

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