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Snow Wetness Influence on Impulse Radar Snow Surveys Theoretical and Laboratory Study

A. Lundberg and H. Thunehed

Luleå University of Technology,
SE-971 87 Luleå, Sweden

The snow-water equivalent of late-winter snowpack is of utmost importance for hydropower production in areas where a large proportion of the reservoir water emanates from snowmelt. Impulse radar can be used to estimate the snow-water equivalent of the snowpack and thus the expected snowmelt discharge. Impulse radar is now in operational use in some Scandinavian basins. With radar technology the radar wave propagation time in the snowpack is converted into snow-water equivalent with help of a parameter usually termed the a-value. Use of radar technology during late winter brings about risk for measurements on wet snow. The a-value for dry snow cannot be used directly for wet snow. We have found that a liquid-water content of 5% (by volume) reduces the a-value by approximately 20%. In this paper an equation, based on snow density and snow liquid water content, for calculation of wet-snow a-value is presented.

Introduction

In Scandinavia a large proportion of the hydropower production emanates from snowmelt in mountainous areas. Hydropower companies need snowmelt runoff forecasts for efficient reservoir operation and estimates of snow-water equivalent (*SWE*) is thus important. In mountains the snow-cover depth is quite variable due to snow redistribution and accurate estimates of *SWE* become difficult. Manual density and snow depth measurements used in traditional snow surveys are costly. Satellite methods are not operational to any large extent in Scandinavia (Lundberg and

Halldin 1999). In order to develop more efficient and accurate methods for snow surveys several efforts to apply impulse radar technology have been made (*e.g.* Ulriksen 1982; 1985; Killingtveit and Sand 1988; Andersen *et al.* 1987; Brandt 1991; Bruland and Sand 1996; Sand and Bruland 1998). Airborne devices are used in Sweden (*e.g.* Ulriksen 1985; Brandt 1991; Lundberg *et al.* 1999) and impulse radar measurements from helicopter have been in operational use in Sweden since 1986 (Wikström 1999, personal communication). Measurements from the snowpack surface are preferred in Norway since the very steep terrain makes it difficult to make airborne measurements (*e.g.* Killingtveit and Sand 1988; Bruland and Sand 1996; Sand and Bruland 1998).

With radar technology, *SWE* is determined from the radar wave propagation time in the snowpack. The wave propagates from a transmitter (located above the snow surface) through the snowpack to the soil surface and then back to a receiver (at the same level as the transmitter). The travel time in air is subtracted. The propagation time in the snowpack, called the two-way travel time (*twt*) is then used to determine *SWE* according to

$$SWE = a\ twt \qquad (twt \text{ in nanoseconds}) \tag{1}$$

Lundberg *et al.* (1999) showed that for dry snow the *a*-value ranges from 0.036 to 0.045m ns^{-1} for densities ranging from 300 to 400 kg m^{-3}. They also showed that a is a function of snow density ρ_S and dry snow dielectric constant ε_{DS}

$$a = \frac{1.5\times10^{-4}\rho_S}{\sqrt{\varepsilon_{DS}}} \qquad (\mathrm{m\,ns^{-1}}) \tag{2}$$

where ε_{DS} can be estimated from ρ_S with

$$\varepsilon_{DS} = (1+0.000851\rho_S)^2 \tag{3}$$

ε_{DS} was calculated with a dielectric mixing model using ε and the volume fractions Θ of the constituents following Roth *et al.* (1990). For dry snow the *a*-value can thus be estimated using snow density only

$$a = \frac{1.5\times10^{-4}\rho_S}{1+0.000851\rho_S} \qquad (\mathrm{m\ ns^{-1}}) \tag{4}$$

Maximum *SWE* is required for optimization of hydropower production. It is thus desirable to make measurements as late as possible during the accumulation season. During late winter melting may occur or the snowpack may be wetted by rain. A day with an average air temperature of 4°C produces melt-water corresponding to approximately 5% liquid water content (by volume) for a snowpack with 0.3 m *SWE*. For practical reasons measurements from aeroplane or helicopter usually have to be scheduled a couple of days ahead and measurements on wet snow may be difficult to avoid. Impulse radar measurements seem to have been restricted to hydropower applications but there are other possible applications. There is a need of ground-truth data for verification of satellite images of *SWE*. Areally averaged *SWE* are also needed for studies of ground-snow-forest-atmospheric exchange (Lundberg and Halldin 1999). In this case *SWE* is needed during the full winter season including autumn and spring. This brings about risk for measurements performed on wet snow.

There is also a need to quantify the magnitude of the errors introduced when measuring on wet snow and to correct for those errors. The effect of liquid water content (θ_W, % by volume) on the *a*-value does not seem to have been comprehensively investigated. Annan *et al.* (1994) sketched the principles for measurements on wet snow but concluded that determination of *SWE* was not possible with ground penetrating radar alone when melt had occurred. Killingtveit and Sand (1988) studied discrepancies ΔSWE between SWE_{RAD} and SWE_{SURV} for wet snow (SWE_{RAD} = *SWE* determined with radar technology, SWE_{SURV} = *SWE* determined with traditional snow surveys). They failed to establish relationships between ΔSWE and ρ_S or liquid water content θ_W. On the other hand there are many reports where the influence of liquid water content on the dielectric constant for wet snow ε_{WS} is measured and modelled. Empirical relationships between ε_{WS} and θ_W and ρ_S are presented by *e.g.* Abdelrazik (1984), Sihvola and Tiuri (1986), Roth *et al.* (1990), Denoth (1989; 94) and Perla (1991). It thus seems likely that it should be possible to derive the wet-snow *a*-value from an expression of ε_{WS}.

Aim

The aim of this study was therefore to derive expressions for the wet snow *a*-value using relationships between ε_{WS} and θ_W. The derived *a*-values were used to calculate *twt* and the calculated *twt*-values were then compared with laboratory radar measurements of *twt*. The comparison was made for snow wetness varying from 0 to 9% (by volume) and for snow densities ranging from 223 to 477 kg m^{-3}.

Experiment Design and Radar Measurements

An experiment was designed to illustrate the effect of θ_W on the measured *twt* for known *SWE* and ρ_S. A plywood box with height 1 m and cross-section area of 0.5 m^2 was filled with dry snow. Fairly new-fallen snow (a couple of days old) was used for the experiments. The initial snow density (kg m^{-3}) was determined by weighing the box empty and full. The box was placed in a temperature controlled room at approximately -2°C and the snow was allowed to adjust to the ambient room temperature.

Radar measurements were conducted using the Raman ground penetrating radar, 450 MHz antenna. The one-way travel time (*owt*) was measured by placing the transmitter antenna on top of the snow and the receiver underneath the plywood box

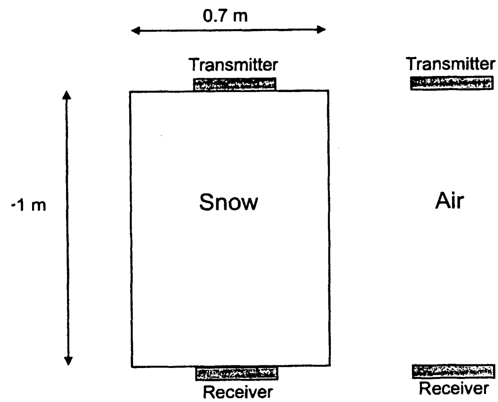

Fig. 1.
Set-up for the laboratory experiment: Transillumination measurements were performed in the snow (left) and compared with the corresponding measurements in air.

(Fig. 1), in order to perform transillumination measurements. A number of traces were recorded in the middle of the box holding the antennas parallel to the sides of the box and the antennas were then removed. The measurements were repeated in air with the same antenna separation distance s. The *owt* in air can be calculated as: $t_A \equiv s/c$, where c is the radar-wave velocity in vacuum 3×10^8 m/s. The *owt* in wet snow was then calculated as: $t_{WS} \equiv t_A + \Delta t$, where Δt was the time delay of the radar wave in snow compared to air (Fig. 2). With this procedure it was not necessary to determine the exact zero time.

The measurements started with dry snow and the θ_W-content was then step-wise increased. Batches of 5-litres of water (chilled to close to zero degrees) were evenly sprinkled over the snowpack with a garden pitcher. The pitcher was equipped with a small-holed nozzle in order to prevent the presence of ice needles in the added water. Before a batch of water was added *owt* was determined with radar measurements. The θ_W was determined by two methods: a combined water and energy budget method and a dilution method. The upper level of the snowpack was measured to keep control of the increase in density due to compaction of the snow-pack. The changes in density due to added water and due to the extraction of snow samples for the dilution measurements were compensated for. Four experiments (a, b, c and d) with varying initial densities were performed. We wanted to start the experiments with snow temperatures just below zero (without risking uncontrolled melt) and water temperatures just above zero (without risking presence of ice needles). The conditions for the experiments listed in Table 1 illustrates that this was not fully achieved. The initial snow temperatures for the experiments were approximately -2°C and the added water temperature was between 0.1 and 0.3°C. The initial densities ranged from 223 to 383 kg m^{-3}. To achieve the high initial snow density of experiment b) the box was step-wise filled with snow. The snow was inserted in approximately 0.2 thick layers and the snow was packed for each inserted layer. Snow wetness was determined with the combined water and energy budget method for all experiments and with the dilution method for experiments a) and b).

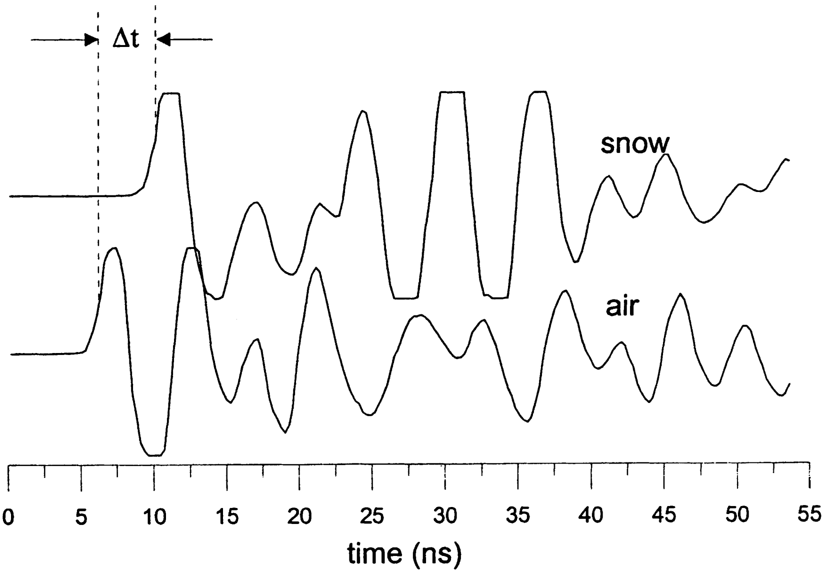

Fig. 2. Example of measured trace in snow (top) and in air (bottom). Only the first arrivals are considered. The position of the time scale is arbitrary since only the difference in arrival time (Δt), *i.e.* the time delay due to the wet snow, is of importance. Δt is in this example 3.9 ns.

Table 1 – Initial and final snowpack densities (ρ_{INIT}, ρ_{FINAL}), heights (h_{INIT}, h_{FINAL}) and temperatures (T_S^{INIT}, T_S^{FINAL}). Final wetness θ_W^{FINAL}), masses (m_W) and average temperatures (T_W) of added water. Mass of water frozen by change in snow cold content (m_W^F). Mass of water melted by heat content of added water (m_W^M).

Variable	Exp. a)	Exp. b)	Exp. c)	Exp. d)
ρ_{INIT} (kg m^{-3})	223	383	256	324
ρ_{FINAL} (kg m^{-3})	251	477	327	427
h_{INIT} (m)	0.95	0.95	0.99	0.95
h_{FINAL} (m)	0.95	0.93	0.96	0.93
θ_W^{FINAL} (% by volume)	2.4	6.0	8.6	9.1
T_S^{INIT} (°C)	-2.27	-2.09	-1.74	-1.96
T_S^{FINAL} (°C)	-0.50	-0.03	- 0.53	0.00
m_W (kg)	13.5	45.6	30.0	45.0
T_W (°C)	0.1	0.3	0.3	0.1
m_W^F (kg)	1.32	2.64	1.08	2.12
m_W^M (kg)	0.051	0.171	0.113	0.056

Wetness Measurements

Water Budget – Energy Budget (WBEB) Method – The liquid water content of a snowpack can be determined from a combined water and energy budget if the temperatures of the snow and added water are known. If the initial snow temperature is below zero the "cold content" of the snowpack will freeze some of the added water. The frozen water mass m_W^F (kg) can be calculated from the snow mass m_S (kg) and the snow temperature T_S (°C) with

$$m_W^F = \left(\frac{m_S C_{ice} T_S}{L}\right) \tag{5}$$

C_{Ice} is the specific heat of ice at 0°C = 2.35×10^3 J kg^-1 °C^-1, and L is the latent heat of fusion, 334×10^3 J kg^-3. The added water will melt some snow if the water temperature is above zero. The melted snow mass m_W^M (kg) can be calculated with Eq. (5) if the snow mass, temperature and specific heat is substituted by corresponding values for the added water, *i.e.* m_W, T_W and $C_W = 4.18\times10^3$ J kg^{-1} °C^{-1}.

Dilution Method – Davis *et al.* (1985) described in detail a method to quickly determine the liquid water fraction x_W of wet snow and we followed their recommendations. Snow samples were taken with a hollow cylinder tube with cross-sectional area 1.96×10^{-3} m^2. A stock solution of mass m_{Stock} and concentration c_{Stock} was mixed thoroughly with the wet snow sample (with mass m_{Sample}) in a thermos. A stock solution of approximately 0.01 N hydrochloric acid and a m_{Stock}/m_{Sample} ratio in the range 0.5 to 0.8 was used. The stock solution was cooled to ≈ 0°C before being added to the wet snow sample in the thermos. The mixture in the insulated container (thermos bottle) was vigorously shaken for at least 30 seconds. A sample of the mixture of the stock solution and the snow liquid water (with concentration c_{Mix}) was then extracted from the thermos with the help of a syringe. The syringe was equipped with a filter (to avoid including ice crystals). The unknown water fraction x_W of the snow sample was then determined as

$$x_w = \frac{m_{Stock}}{m_{Sample}} \left[\frac{1 - \frac{C_{mix}}{C_{Stock}}}{\frac{C_{Mix}}{C_{Stock}} - \frac{C_{Water}}{C_{Stock}}} - \frac{T_{Stock} C_W}{L} \right] \tag{6}$$

where c_{Water} is the concentration of the liquid water in the snow. The conductivity was determined with a CDM 83 Conductivity meter (Radiometer, Copenhagen) and then transformed to concentrations as shown in Davis *et al.* (1985). The last term, containing the temperature of the added stock solution T_{Stock}, C_W and L compensated for the snow melted by the stock solution (if the solution had a temperature above 0°C). The volumetric liquid water content θ_W was then determined by

$$\theta_W = \frac{100 x_W \rho_S}{\rho_W} \tag{7}$$

where ρ_W is the density of water.

Theory

The a-value for wet snow can be calculated with Eq. (2) if ε_{DS} is substituted by the dielectric constant for the wet snow ε_{WS} (Lundberg *et al.* 1999).

$$a = \frac{1.5\times10^{-4}\rho_S}{\sqrt{\varepsilon_{WS}}} \qquad (\mathrm{m\ ns}^{-1}) \tag{8}$$

Dielectric Properties of Wet Snow

The dielectric constant of a mixed non-magnetic media can be estimated with a three phase mixing model (*e.g.* Birchak *et al.* 1974; Roth *et al.* 1990; Perla 1991)

$$\varepsilon_{WS} = (\theta_W \varepsilon_W^q + \theta_1 \varepsilon_1^q + \theta_A \varepsilon_A)^{1/q} \tag{9}$$

where θ are the volume fractions of the constituents. Here $_{W,I,A}$ denotes the constituents of wet snow *i.e.* water, ice and air respectively. Many authors (*e.g.* Freedman and Vogiatzis 1979; Roth *et al.* 1990; Bergström 1997) have found good agreement between calculated and measured ε using the three phase mixing model and $q \approx 0.5$ for different geological media. For wet snow Lundberg (1997) found fair agreement between measured and modelled ε with the same q -value. The q-values optimized by Perla (1991) for wet snow range around 0.5 (See Table 2). With $q = 0.5$ the denominator of Eq. (8) becomes

$$\sqrt{\varepsilon_{WS}} = \theta_W\sqrt{\varepsilon_W} + \theta_1\sqrt{\varepsilon_1} + \theta_A\sqrt{\varepsilon_A} \tag{10}$$

Eq. (10) can be expanded and simplified into

$$\sqrt{\varepsilon_{WS}} = 1 + c_1\rho_S + c_2\theta_W \tag{11}$$

The simplification is made using the relationship between the volume constituents, $\theta_A = 1 - \theta_I - \theta_W$ and the definition of $\rho_S = (\theta_W\rho_W + \theta_I\rho_I + \theta_A\rho_A)$ (Lundberg *et al.*

Table 2 – Optimised q-values for different snow wetness θ_W (Modified from Perla 1991).

θ_W (% by volume)	q
$\theta \le 3$	0.625 ± 0.065
$3 < \theta \le 50$	0.43 ± 0.02
$50 < \theta < 100$	0.51 ± 0.01

1999). The fact that $\rho_A << \rho_W$ and ρ_I was also used. The ε_W is temperature and frequency dependent. The value for ε_W varies from about 88 at 0°C to 81 at 20°C and can be regarded as independent of frequency for frequencies < 1 GHz (Kendra *et al.* 1994; Perla 1991). For numerical values of $\varepsilon_W = 88$, $\varepsilon_I = 3.17$, $\varepsilon_A = 1.0$ and $\rho_I = 917$ kg m^{-3} the constants c_1 and c_2 became 0.000851 and 7.529 respectively (Lundberg *et al.* 1999). Eqs. (8) and (11) gives

$$a = \frac{1.5\times10^{-4}\rho_S}{1+c_1\rho_S+c_2\theta_W} \qquad (\mathrm{m\ ns}^{-1}) \tag{12}$$

Denoth (1989; 94) and Sihvola and Tiuri (1986) separated the influence of ρ_S and θ_W on ε_{WS} according to

$$\varepsilon_{WS} = 1 + \underbrace{c_3\rho_S + c_4\rho_S^2}_{\rho_S\ \text{variation}} + \underbrace{c_5\theta_W + c_6\theta_W^2}_{\theta_W\ \text{variation}} \tag{13}$$

A similar formula can be derived from Eq. (11)

$$\varepsilon_{WS} = \left((1+c_1\rho_S)+c_2\theta_W\right)^2 = (1+c_1\rho_S)^2 + 2(1+c_1\rho_S)c_2\theta_W + (c_2\theta_W)^2 \tag{14}$$

$$\varepsilon_{WS} = 1 + 2c_1\rho_S + c_1^2\rho_S^2 + 2c_2\theta_W + c_2^2\theta_W^2 + 2c_1c_2\rho_S\theta_W \tag{15}$$

Eq. (15) is similar to Eq. (13) with $c_3 = 2c_1$, $c_4 = c_1^2$, $c_5 = 2c_2$, $c_6 = c_2^2$. An additional term with the constant $c_7 = 2c_1c_2$ in front of the product $\rho_s\theta_W$ also emerges. The constants in Eq. (13) calculated from the mixing model using $q = 0.5$ are shown in Table 3 together with constants determined empirically by Sihvola and Tiuri (1986) and Denoth (1989; 94). The dielectric constants for pure ice ε_I and for pure water ε_W were also calculated (See last two columns of Table 3 using Eq. (13) and the con-

Table 3 – Constants in Eq. (13) from different studies. Estimates of ε_I and ε_W calculated using Eq. (13) and the constants.

Study	c_3	c_4	c_5	c_6	c_7	frequency	ε_W	ε_I
D	1.92×10^{-3}	4.4×10^{-7}	18.7	45.0	-	20 MHz	67.1	3.14
S	1.7×10^{-3}	7.0×10^{-7}	8.7	70.0	-	1 GHz	81.9	3.16
R	1.7×10^{-3}	7.244×10^{-7}	15.06	56.7	0.0128	<1 GHz	88.0	3.18
*	1.7×10^{-3}	7.244×10^{-7}	23.7	56.7	0.0128	<1 GHz	96.6	3.18
P						1MHz	88.0	3.18

D = Denoth (1989; 94), *S* = Sihvola and Tiuri (1986), *R* = Derived from Roth *et al.* (1990) assuming $e_W = 88$ *i.e.* frequency <1 GHz , * Modified from Roth to fit measured data, *P* = Perla *et a*l. (1991) ε_W calculated with Eq. (15) and $\theta_W = 100\%$.

stants c_3-c_7 from Table 3. A combination of Eqs. (1), (8) and (15) gives the *twt*-value

$$twt = \frac{SWE\sqrt{1+c_3\rho_S + c_4\rho_S^2 + c_5\theta_W + c_6\theta_W + c_7\theta_W\rho_S}}{1.5\times10^{-4}\rho} \tag{16}$$

A combination of Eqs. (1), (8) and (11) gives a simpler expression for *twt*

$$twt = \frac{SWE(1+c_1\rho_S + c_2\theta_W)}{1.5\times10^{-4}\rho_S} \tag{17}$$

The *a*-value for wet snow can thus be calculated

$$a = \frac{1.5\times10^{-4}\rho_S}{1+0.000851\rho_S + 7.529\theta_W} \quad (\mathrm{m\ ns}^{-1}) \tag{18}$$

Another empirical method to estimate ε for a composite media used on snow (Perla 1991) is

$$\varepsilon_{WS} \equiv \sum_{i=I,A,W} \varepsilon_i\theta_i - 0.43\theta_A\theta_I - 50\theta_W\theta_I - 100\theta_A\theta_W \tag{19}$$

The constants (-0.43, -50 and -100) given in Eq. (19) were determined by Perla (1991) using the full range of ρ_W and ρ_S. Combination of Eqs. (1), (8) and (19) then gives the *twt*-value

$$twt = SWE\,\frac{\sqrt{\varepsilon_{WS} = \sum_{i=I,A,Q} \varepsilon_i\theta_i - 43\theta_A\theta_I - 50\theta_W\theta_I - 100\theta_A\theta_W}}{1.5\times10^{-4}\rho_S} \tag{20}$$

Comparison between Models and Measurements

The *owt*-values determined with radar measurements were converted to *twt*-values. The *twt*-values were modelled using Eq. (16) and the values for the constants listed in Table 3. Constants derived from Roth *et al.* (1990) are denoted (*R*) and constants from Denoth (1989; 1994) and Shivola and Tiuri (1986) are denoted (*D*) and (*S*) respectively. Note that Eq. (17) is a special case of Eq. (16) with the constants given by (*R*). *twt*-values were also determined with Eq. (20). The modelled *twt*-values were compared with the *twt*-values determined with radar technique.

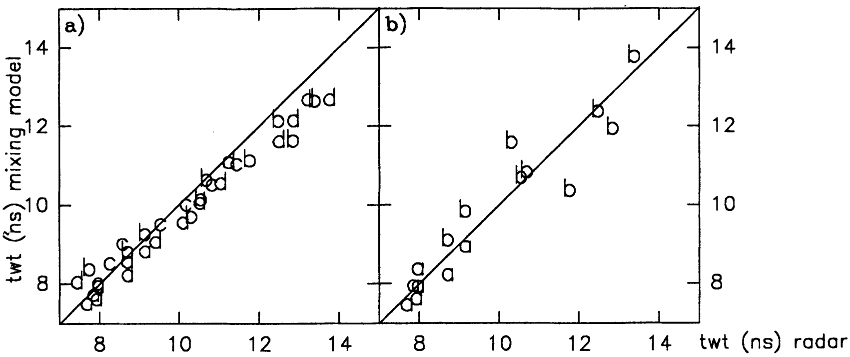

Fig. 3. *twt*-values modelled with Eq. (17) (using constants R in Table 3) compared with *twt*-values determined by radar measurements. The letters denote the different experiments listed in Table 1. a) Wetness determined with WBEB. b) Wetness determined with dilution method.

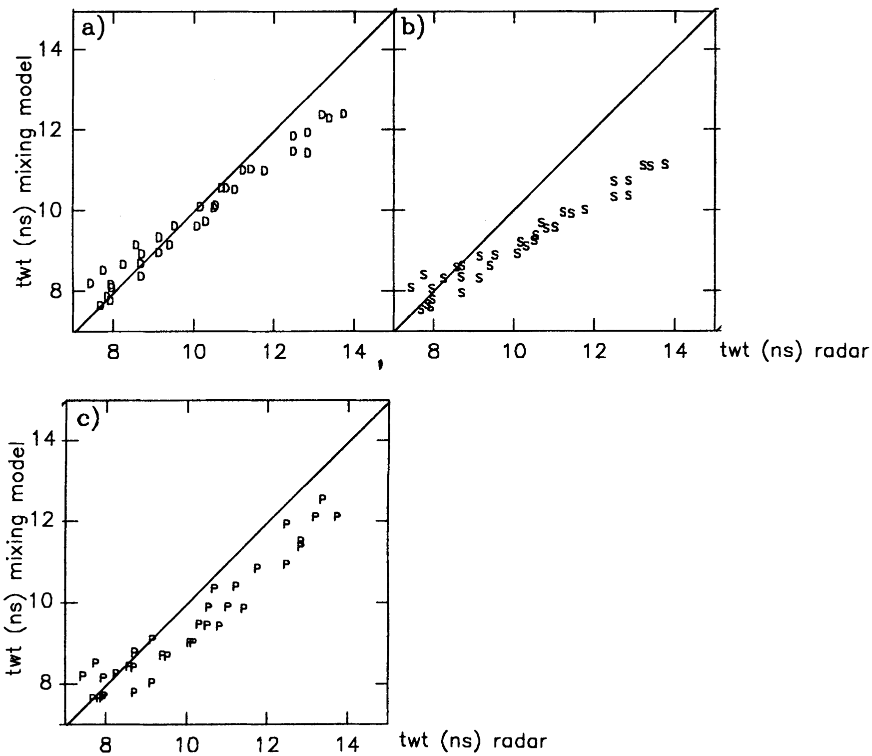

Fig. 4. Modelled *twt*-values (WBEB method used to determine wetness) versus *twt*-values measured with radar technique. Modelled *twt*-values using Eq. (16) and constants from Table 3. a) D constants from Denoth (1989; 94), b) S constants from Shivola and Tiuri (1986), c) *Twt* modelled with Eq. (19) derived using ε_{WS} from Perla (1991).

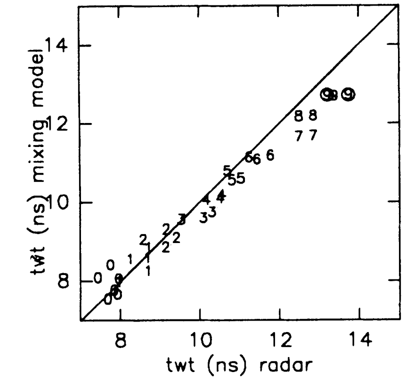

Fig. 5. Modelled *twt*-values using Eq. (17) versus *twt*-values determined with radar technology. The figures denote approximate snow wetness (% by volume). Wetness determined with WBEB-method.

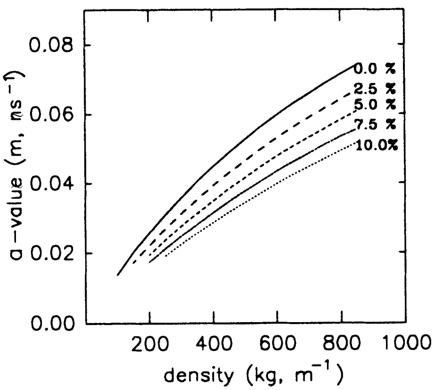

Fig. 6. The *a*-value in Eq. (18) as function of snow density ρ_S and snow liquid water content θ_W. Wetness determined with WBEB-method.

Results

The *twt*-values determined using the constants (*R*) from Table 3 and Eq. (17) agree fairly well with the *twt*-values determined with radar measurements (Fig. 3). The agreement between modelled and measured *twt*-values was slightly better when the wetness was determined with the WBEB method than with the dilution method (*cf.* Figs. 3a and b). The model predicts slightly lower *twt*-values than the measured *twt*-values for high *twt*-values (Fig. 3a). The constants (*R*) (*cf.* Figs. 3a and 4a) give a slightly better fit to the measured data than the constants (*D*). The constants (*S*) and also using ε_{WS} from Perla (1991) for modelling *twt* (Eq. 19) seem to underestimate the *twt*-values (see Figs. 4 b and c). The agreement between the measured and modelled *twt*-values using Eq. (17) is good for θ_W less than 6% (Fig. 5) while for higher θ_W-values the modelled *twt*-values are slightly lower than the measured values. The two circled measurements with $\theta_W = 9\%$ were measured before and after a 90° change of the orientation of the transmitting and the receiving antennae. The *a*-values calculated with Eq. (18) for different θ_W illustrates that the *a*-value is approximately 20% lower for a snow with θ_W=5% compared to dry snow (Fig. 6 and Table 4).

Table 4 – The a-value in Eq. (18) as function of snow density ρ_S and snow liquid water content θ_W (% by volume).

ρ_S (kg m^{-3})	θ_W (%) 0.0	2.5	5.0	7.5	10.0	$\frac{a_{0\%}-a_{5\%}}{a_{0\%}}$
100	0.014					
200	0.026	0.022	0.019	0.017	0.016	24
300	0.036	0.031	0.028	0.025	0.022	23
400	0.045	0.039	0.035	0.032	0.029	22
500	0.053	0.046	0.042	0.038	0.034	21
600	0.060	0.053	0.048	0.043	0.040	20
700	0.066	0.059	0.053	0.049	0.045	19
800	0.071	0.064	0.058	0.053	0.049	18
900	0.076	0.069	0.063	0.058	0.054	18

Discussion

The modelled *twt*-values agree acceptably well with the measured *twt*-values for the constants (*R)* and (*D*), even if the modelled values are a bit underestimated at high wetness. A better fit between modelled and measured *twt*-values can easily be achieved by increasing one of the constants in front of θ_W in Table 3 (*i.e.* constants c_5-c_7). If *e.g.* the (*R*)-constants are used but c_5 is increased to 23.7 instead of 15.05 a better fit between modelled *twt*-values will be achieved. The ε_W-value calculated with those constants (θ_W = 100%) will however become 96.6 which is far from the correct value 88.0 (see * in Table 3). Also the (*D*)-constants do not give a realistic value of ε_W (see Table 3).

The temperature in the environmental room was set to -1°C during the experiments, but the temperature fluctuated a few degrees. It is not likely that these fluctuations could have produced melt or refreezing to such an extent that corrections should be made. The heat content of the added water (corresponding to a maximum melt of less than 0.2 kg or 0.1% θ_W) was not compensated for in the calculations. Nor was the "cold content« of the cold air in the snowpack considered.

The dilution method was expected to give more accurate θ_W -values than the WBEB method and dilution measurements were planned for all experiments. During experiment b), with packed snow, we noticed that the wetness of the snow samples did not seem to increase evenly when we added more water. The sizes of the snow samples were too small to give accurate average θ_W-values when the snow was packed. It was not possible to increase the number of samples without significantly disturbing the snow pack and the dilution method was thus abandoned. The infiltrated water was probably not evenly distributed and denser snow layers prevented the water from percolating evenly through the snow pack and preferred pathways for liquid H_2O was probably developed. The discrepancy between the two radar measurements (Δtwt (1 ns) made for the same wetness and density, but with a 90 degree change in orientation for the radar measurements, also illustrates the anisotropy of the snow wetness (Fig. 5). At the highest liquid water contents the water was concentrated at the base of the snowpack. We did not verify that the base of the box was horizontal during the experiments neither did we assure that no depression of the box base occurred. It was thus possible that more water could accumulate at some locations than at others at the base of the snow pack. This could account for at least part of the discrepancy between measured and modelled *twt*-values at high θ_W contents.

A possible difficulty with radar measurements on wet snow in the field, (not treated here) is the difficulty to distinguish the reflection in a nearly saturated layer at the base of the snowpack from the reflection at the ground surface. This difficulty was avoided by measuring *otw* instead of *twt* in our experiments.

Snow Wetness Estimates

It is advisable to avoid making snow radar measurements when the snow is wet since it is difficult and time consuming to measure θ_W. Comparative studies for θ_W measurement techniques and instruments are presented by *e.g.* Denoth *et al.* (1984) and Boyne and Fisk (1987). Methods for *in situ* determination of θ_W are listed in Table 5. An indirect method to estimate snow wetness resulting from melt is the so-called degree-day or index method. The daily melt Q_{MELT} (mm, day^{-1}) is calculated as the product of a degree-day-factor DF (mm,°C^{-1}, day^{-1}) and the number of degrees the average daily air-temperature T (°C) exceeds a threshold temperature T_T (°C)

$$Q_{MELT} = DF(T-T_T) \tag{21}$$

DF and T_T determined for rural areas in northern Sweden and Alaska are shown in Table 6. Both studies using watershed runoff techniques use DF = 3.5 (mm,°C^{-1}, day^{-1}) while the DF-values determined using snow courses are more spread. The degree-index method does not include radiation and when the melt periods were separated into early and late melt the DF became lower for early melt than for late melt (See Table 6).

Table 5 – Methods for *in situ* measurements of snow wetness.

Method	Reference	Method	Reference
Calorimetric method	Radok *et al.* (1961)	TDR-method	Schneebeli and Coléou (1997)
Dilution method	Davis *et al.* (1985)	Monopole-antenna	Denoth (1997)
Snow-fork	Sihvola and Tiuri (1986)	Capacitance sounding	Louge *et al.* (1998)

Table 6 – Degree-day-factors DF (separated into early and late snowmelt period) and threshold temperatures T_T (C) determined for open fields in northern Sweden and Alaska.

Study	Technique	Location	DF (mm,C^{-1}, day^{-1})			T_T (C)
			average	early	late	
Bergström (1990)	Watershed runoff	Northern Sweden	≈ 3.5			0
Hinzeman & Kane (1991)	Watershed runoff	Alaska, 68°37'N 149°17'W	3.5			-1.9 to 0.5
Lundberg, (1979)	Snow courses	Luleå, 65°38'N 22°15'O	6.1	3.0[1]	5.9 – 8.6[2]	0
Westerström, (1982)	Large lysimeters	Luleå, 65°38'N 22°15'O		0.7 – 1.7	3.3 – 6.5	0
Beyerel (1999)	Snow courses	Luleå, 65°38'N 22°15'O	3.3	1.0[3]	5.7[4]	0
Kane *et al.* (1997)	Snow courses	Alaska, 68°37'N 149°17'W	2.7			-0.2

[1] 15-24 April, [2] 25 April-10 May, [3] 30 March-15 April, [4] 16 April-23 April

The order of magnitude of θ_W produced by melt on open fields can be estimated using the *DF*-value (3.5 mm,°C-1, day-1) and T_T (≈0°C) determined from runoff studies. During a 24-hour period with an average air temperature of 4°C a melt of ≈ 14 mm is produced. This melt corresponds to θ_W ≈5% for a snowpack with 1 m depth and a density of 300 kg/m^3.

Melt takes place at the snow surface, with free air access. Refreezing takes place throughout the entire snowpack and the air has to penetrate the whole snowpack in order to refreeze the melted water. Air-temperature is thus less effective when it comes to refreezing the melted water. An order of magnitude smaller *DF* is usually used for refreezing. This means that it will require 10 days with an average air-temperature of -4°C to assure that all the melted water in this example has refrozen.

If θ_W and ρ_S can be regarded as fairly constant over the studied area the *a*-value can be determined empirically by using results from the simultaneous measurements of *twt* and *SWE* at the same location. Another possibility might be to make radar velocity measurements from the snow surface using common midpoint technique (CMP) or wide-angle reflection and refraction (WARR) technique (*e.g.* Parasnis 1997) and thus determine ε_{WS}.

Conclusion

The *a*-value calculated from snow density only should not be used for wet snow. The actual wet-snow *a*-value is lower than the dry-snow value. A liquid-water content of 5% (by volume) reduces the *a*-value by ≈20%. Snow radar measurements should be avoided during periods with liquid water in the snowpack. If this is not possible, measurements or estimates of θ_W should be made and the *a*-value should be corrected for θ_W. Eqs. (17) and (18), derived from the mixing model (Eq. (9)) with $q = 0.5$ give:

- good agreement between modelled and measured *twt*-values for the normal wetness range for old snow ($\theta_W \leq 4\%$)
- fair agreement between modelled and measured *twt*-values for extremely wet snow ($6\% \leq \theta_W \leq 9\%$)
- a simple expression for $\sqrt{\varepsilon_{WS}}$
- a correct estimate of ε_W
- simple expressions for *twt* and the *a*-value.

The equations thus seem well suited for estimates of snow wetness influence on impulse radar measurements. It would be helpful if field measurements were available to verify the findings.

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Address:
Angela Lundberg,
Water Resources Engineering,
Luleå University of Technology,
SE-971 87 Luleå,
Sweden.

Email: angela.lundberg@sb.luth.se

H.Thunehed,
Applied Geophysics,
Luleå University of Technology,
SE-971 87 Luleå,
Sweden.

Email: Hans.Thunehed@sb.luth.se