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1. Introduction

The purpose of this document is to provide the theoretical basis for the estimation of snow depth by means of satellite altimeter data. The method here proposed is mostly oriented for the CryoSat-2 SIRAL instrument, however it could potentially be used in any kind of radar altimeter, provided that the data product allow to compute a calibrated value of the surface backscattering coefficient.

The following section provides a description of the most important snow parameters, and the interaction of snow with electromagnetic waves, which provides the physical foundations for the estimation of snow depth. Section 3, shows previous results for snow depth estimation with radar altimeters, and finally, Section 4 gathers the methodology description of the algorithm for snow depth estimation.

This algorithm has been applied to SAR and SARIn data of the CryoSat SIRAL instrument. The results and the comparison with snow depth ground truth are gathered in D2.6: *Develop Processing for Snow Depth.*

2. Remote Sensing of Snow Parameters

The study of snow and the characterization of snow pack evolution is a key parameter in general, but in particular for regions where water supply is mainly due to snow melt and runoff.

The importance of the estimation of snow parameters, such as the Snow Water Equivalent (SWE), which quantifies the amount of water contained in the snow-pack, has pushed the scientific community to explore the possibilities of the Earth Observation techniques for the snow quantitative estimation. SWE is one of the most important measurements of a snow-pack from a hydrological point of view because when some, or all, of the water contained in the snow is released during the seasonal melt, it becomes an important component of local surface water and groundwater budgets.

Snow Water Equivalent (SWE) represents the amount of water contained within the snow-pack. In practise, it is equivalent to the quantity of water that would be obtained if you melted the entire snow- pack. This parameter is further used in hydrology and climate studies for the forecasting of water resources, natural diseases such as flooding, and water resources management in the scientific field, but it is also important for other activities such as hydro-power generation.

In order to estimate the SWE of a snow-pack, two parameters have to be estimated: the snow depth (d) and the snow density (ρ) in order to obtain the SWE estimation as the product of them. However, SWE measurement is tedious and costly, since it consists of melting a known volume of snow in order to weight the water contained in it. In field measurement campaigns, that cannot be carried out in vast areas. Therefore SWE cannot be extensively studied for large scale applications which are the ones of operational interest. From this need of mapping in higher spatial and frequency scales the SWE parameter, arises the interest of the remote sensing community in the estimation of this snow-pack parameter.

The remote sensing for snow monitoring has been a research topic for the last three decades. The snow monitoring help us to have an adequate understanding of snow and the water cycle. According to Mätzler (1987), monitoring the amount and distribution of snow, helps in the climate study, such as long-term changes in the climate system and their potential impacts on human life, the management of clean water resources, improvements of the weather forecasts, as well as water supply and flooding predictions for hydrology management.

Some studies have demonstrated the ability to calculate the water equivalent of dry snow with the use of a technique called "InSAR" (Interferometric Synthetic Aperture Radar). This technique is based on the measure of the phase delay introduced in the SAR signal due to the presence of snow on the ground. In case of dry snow coverage, the radar backscattering is from the snow ground interface.

On the other hand, radar altimeters are able to measure the height of the snow pack, due to the effect induced in the backscattering coefficient by the presence of snow. The following sections provide the theoretical basis for the estimation of snow depth from radar altimeters.

2.1. Snow interaction with EM waves

This section starts with a general introduction on the interaction of snow with EM waves in order to understand the main effects of using an active sensor for snow monitoring. Then the components of the snow-pack (ice, air and water) are detailed since the obtained backscattering depends on their variations.

In radar distributed targets, the scattering coefficient σ, characterises the scattering radiation being imaged by the radar. The value of this scattering coefficient depends, on one side, on a set of parameters that belong to the imaging system: wave frequency, wave polarization and image configuration (incident and scattered directions). On the other side, $\sigma 0$ depends on the target characteristics, such as the geometrical or the dielectric properties. Remote sensing techniques aim to extract information about the imaged target out of the retrieved scattered electromagnetic waves. The idea is to have a well modelled environment so that the influence of the target specific factors can be extracted from the gathered scattered power, and therefore information about the target can be retrieved.

In the case of the snow-pack, according to what is detailed in Mätzler (1987) and Ulaby et al. (1986), snow is considerably well modelled by a cloud of ice particles in an air background. However, the interfaces air-snow and snow-ground add some complexity to the method to infer the backscattered power, since it results in a combination of several scattering phenomena.

The backscattering response of the snow-pack is the result of the contribution of different scattering mechanisms: underlying ground surface back-scattering, volume scattering from the snow layer, air-snow interface back-scattering, and multiple contributions from double and triple bounce between the snow-pack and the soil surface. In terms of microwave interactions with the snow-pack, the driving electromagnetic parameters are the relative dielectric constants of ice and liquid water and their geometrical distribution.

2.2. Snow physical modelling

The snow back-scattering depends on parameters such as: density, which can be related to the dielectric constant, the water content, the temperature, which can be related to the liquid water content, and the crystalline structure of the ice particles present in the snow-pack. Snow is composed of three elements: ice, air and water, which can be liquid or vapour depending on the kind of snow, and the variation of these three component percentages, determines the snow features.

The parameters that determine the nature of each kind of snow are introduced as follows:

Water Content (liquid): Determines the dielectric properties and it is very attached to the temperature. Snow is considered dry if there is 0% of liquid water content, and as the temperature raises, the amount of liquid water content increases leading to the so called wet snow (Ulaby et al., 1986).

Density: Snow density is defined as the ratio of snow mass over the water reference mass. It's the most important parameter influencing the dry snow backscattering power. Snow density can be directly related with the dielectric constant with the semi-empirical Looyenga's formula [Singh and Venkataraman, 2008]:

$$\epsilon_s = 1 + 1.5995\rho + 1.861\rho^3 \quad (1)$$

Dielectric constant: The snow dielectric constant real and imaginary parts are obtained from the air and the ice dielectric constant, since snow is an inhomogeneous air and ice mixture. For dry snow, the dielectric constant is almost independent of the frequency and the temperature at the microwaves band of the spectrum, and the imaginary component is negligible whereas the real one is a function of the density. In the wet snow case, the imaginary component of the dielectric constant is strongly related with the liquid water content of snow, increasing rapidly with the presence of liquid water in the snow-pack. The complex dielectric constant is a weighted average of the dielectric constants of the snow components: air, ice and liquid water

$$\epsilon = \epsilon' + i\epsilon'' \quad (2)$$

The real part of the dielectric constant ε' is strongly dependent on the snow density, specially for dry snow. As a summary idea, the more liquid water content, the higher the dielectric constant, and therefore bigger absorption which means less penetration.

Penetration: It depends on the frequency of the incident electromagnetic wave as well as on the dielectric constant of the snow, which is driven by the amount of liquid water. The lower the frequency and the amount of liquid water, the higher the penetration. Penetration is a very important parameter for the remote sensing of the SWE, since the penetration capability of the incident electromagnetic waves will determine the potential retrieve of information of the back-scattered measured waves. In Figure 2.1, the penetration is given as a function of the liquid water content of the snow, in the X axis, for several bands

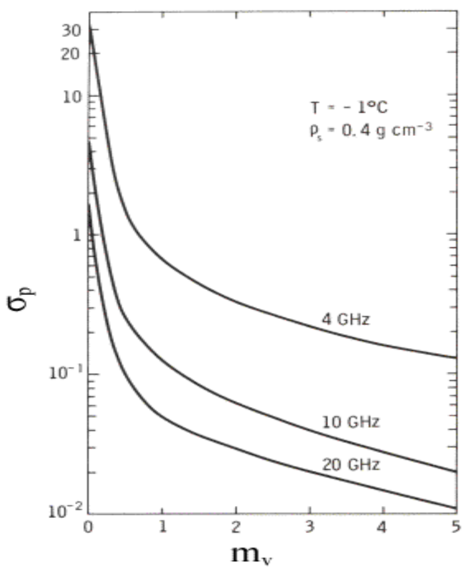

Figure 2.1 Snow penetration depth vs frequency

2.3. Snow-pack Microwaves Scattering Signature

Snow-pack is a volume of ice particles and therefore volume scattering has to be addressed when illuminating snow with EM waves. Volume scattering is caused mainly by the dielectric discontinuities within a volume, whose spatial location is usually random, and then scattering is expected in all directions (Ulaby et al., 1986). The backscattered power received at the radar is the result of the addition of all the volume scatterers contribution within the solid angle defined by the antennas radiation pattern (Mätzler, 1987).

Snow is a mixture of air and ice and its permittivity is a function of the snow density ρ and the relative permittivities of air and ice. Snow is classified in two types depending on the amount of liquid water content per unit volume, since its characteristics and the scattering signature of the snow-pack strongly depend on it.

Dry snow, which has no liquid water content, has a small absorption coefficient allowing microwaves to propagate over long distances, having strong scattering effects in the interaction with the snow-pack volume. The real part of the permittivity ε' of dry snow only depends on snow density which in natural conditions ranges (Ulaby et al., 1986): $0.2 < \rho < 0.5\ g/cm^3$

On the other hand, wet snow, with presence of liquid water content, has a strong absorption coefficient attenuating the microwaves in a very short distance. In such a situation, scattering is limited to a thin layer close to the surface called surface scattering. This is due to the strong dependence of the imaginary part of the permittivity ε'' with the amount of liquid water in the snow volume. Indeed, it can raise an order of magnitude due to an increase of a 0.5 % of the liquid water volume, drastically reducing the penetration depth (Ulaby et al., 1986).

Another interesting analysis involves the incidence angle in a mono-static radar configuration:

- For small incidence angles (close to nadir) surface scattering is the dominating phenomena.
- For bigger incidence angles volume scattering becomes more important, since the surface contribution is moving to forward scattering.

As already introduced, the back-scattering response of the snow-pack is the result of the contribution of different scattering phenomena: underlying ground surface back-scattering, volume back-scattering from the snow layer, air-snow interface back-scattering, and multiple contributions from double and triple bounce between the snow-pack and the soil surface. In Figure 2.2 from [Martini, 2005], gives a clear graphical idea of the multiple scattering mechanisms that compose snow-pack back-scattering.

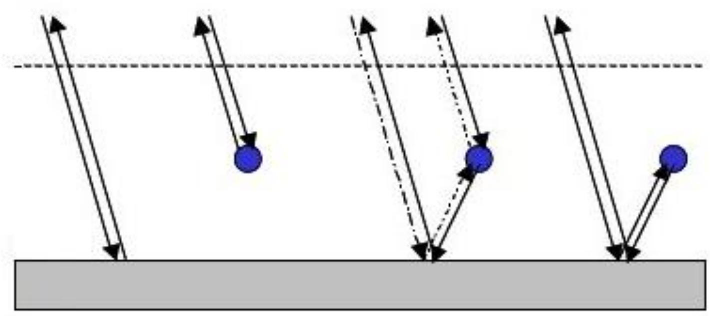

Figure 2.2: Snow scattering mechanisms, [Martini, 2005]

The low dielectric contrast between air and dry snow-pack and the smoothness of the snow surface lead to a snow surface back-scattering coefficient negligible (Ulaby et al., 1986 and Mätzler, 1987), and therefore the scattering signature of the dry snow pack is the addition of the volume scattering contribution and the soil surface back-scattering contribution.

3. Snow Depth Estimation with Altimeters

Altimetry has been used successfully over ice caps, where the surface topography is relatively smooth, for the retrieval of snow depth. Some attempts have been also done over solid earth with relatively flat topography (Papa et al., 2002). The authors used data acquired by the Topex-Poseidon altimeter over the Northern Great Plains, a site already used for numerous experiments with microwaves passive data, due to its flat topography and the presence of more than 280 stations that report the snow depth on a daily basis.

The analysis was conducted over the different parameters that is possible to obtain from altimeter measurements (altimeter range, backscatter coefficient, leading edge width, and trailing edge slope). The most sensitive to SD proved to be the backscattering coefficient, which resulted to decrease during the winter as the snow cover increased along the altimeter track.

This can be explained by the fact that in presence of dry snow, the backscattering coefficient is the result of three different contributions: reflection at the snow/air interface, snow volume scattering and the reflection at the ground/snow interface attenuated twice through the snow layer. Thus increases of snow depth leads to a stronger attenuation of the backscattered signal.

Based on these findings the authors developed a model to retrieve snow depth from single frequency altimeter data. The formulation is based on the estimation of the extinction coefficient which is computed relating the data acquired in presence of snow with snow-free ones.

This methodology has been applied during a project (lead by Starlab), investigating EO capability for snow pack characterisation. Specifically, a work done by IsardSAT, has highlighted some limitation of this technique when applied in mountainous areas, due to the high topography variability within the footprint of the instrument.

As an example in Figure 3.1, from [Reppucci, Escorihuela, 2011], the backscatter temporal signature from Envisat RA is shown over one pixel over the North Great plains (top), which shows a clear seasonal pattern, and the backscatter temporal signature over one pixel in the Pyrenees (bottom), which shows no pattern.

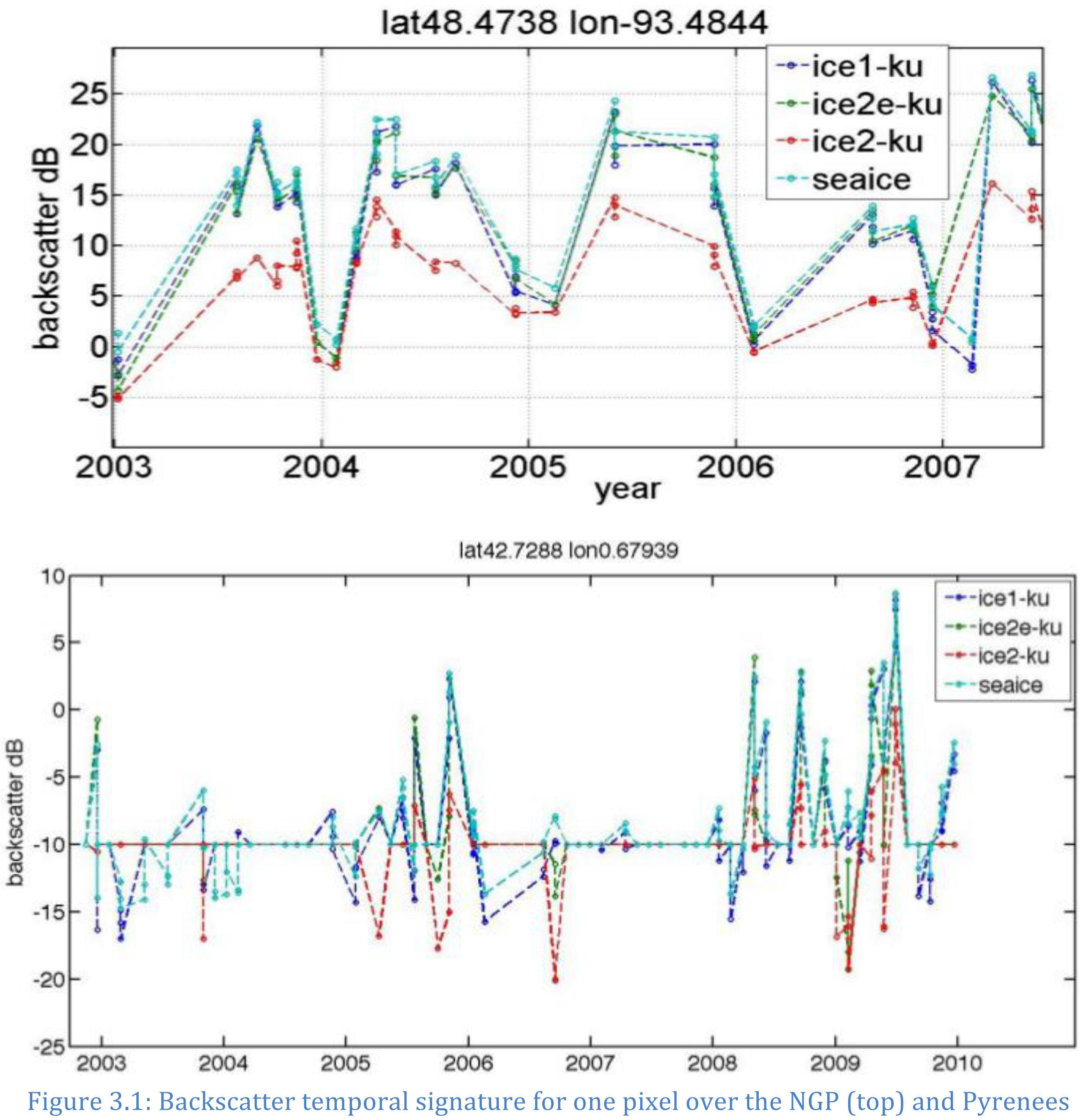

Figure 3.1: Backscatter temporal signature for one pixel over the NGP (top) and Pyrenees (bottom), isardSAT©, from [Reppucci, Escorihuela, 2011].

4. Snow Depth Estimation Methodology Description

4.1. Snow Depth Observables

As explained in the previous section, the main parameter that is expected to be sensitive to snow depth changes is the backscattering coefficient. Thus the proposed methodology will be based on the analysis of this variable.

Following [Papa et al. 2002], the estimation of snow depth will be based on a change detection technique. In details, we can model the backscattering in absence of snow (σ_{ref}) as:

$$\sigma_{ref} = 10 \ln \left(\sigma_{ground}\right) \tag{3}$$

and the backscattering from a snow covered surface as the sum of the snow surface contribution ($\sigma_{surface}$), and two-way attenuation of the ground return signal (σ_{ground}):

$$\sigma_{tot} = 10 \ln \left(\sigma_{ground} * \exp(-2\ k_e\, h) + \ \sigma_{surface}\right) \tag{4}$$

where k_e is extinction coefficient and h the snow depth. Thus combining equation (3) and (4) we can monitor the temporal evolution of the snow pack. The ground backscattering information will be obtained out of the SAR-Altimeter waveforms as explained in the following section.

As the repeat cycle of SIRAL is about 365 days there is almost no overlap of the data samples through a snow season. Thus we had to leave the idea of a “repeat pass approach” and look for neighbouring points, within a given spatial and temporal threshold.

4.2. Backscattering Information Retrieval

As exposed above, the estimation of the snow depth is based on the retrieval of the surface backscattering from the radar altimeter waveform. The challenge is to be able to obtain a consistent measurement of the waveform power in order to relate it to the transmitted power and the rest of system parameters. This will ultimately allow obtaining the Normalized Radar Cross Section, or sigma nought (σ^0), of the surface.

The first step is to convert the altimeter waveform in absolute power. The power echo simple values are scaled to fit between 0 and 65535. The scaling factors can changed for each waveform. To convert these back to values in Watts the following equation should be used, as specified in the CryoSat Product Handbook:

Power in Watts = scaled value * (scale factor * 10^-9) * 2^scale power

Both the scale factor and scale power are provided within the Level-1b CryoSat data product. Once the waveforms have been re-scaled, absolute calibrated σ0 measurements can be obtained out of the waveform power value at the waveform retracking stage, as shown in [Dinardo, 2013].

The sigma nought in dB is derived from the estimated waveform power in Watts, P_u, by inverting the SAR Radar Link Equation. The link is between the reflection surface and the antenna flange. Hence, the SAR Radar Link Equation is given by:

$$\sigma_0 = 10 \cdot \log 10\left(\frac{P_u}{Tx\_Pwr}\right) + 10 \cdot \log 10(K) + bias\_sigma\_0 \tag{5}$$

where:

$$K = \frac{(4\pi)^3 \cdot R^4 \cdot L_{atm} \cdot L_{RF}}{\lambda_0^2 \cdot G_0^2 \cdot A_{SAR}} \tag{6}$$

where A_{SAR} is the resolution ground-cell in SAR-mode:

$$A_{SAR} = (2 \cdot L_y) \cdot (wf \cdot L_X) \cdot 0.886 \cdot \alpha_{Earth} \tag{7}$$

with:

$$\begin{cases} L_y = \sqrt{\dfrac{c_0 \cdot R \cdot PTR\_width}{\alpha_{Earth}}} \\ L_x = \dfrac{\lambda_0 \cdot R}{2 \cdot V_s \cdot \tau_B} \end{cases} \tag{8}$$

And

$$\alpha_{Earth} = 1 + \frac{R}{R_\oplus} \tag{9}$$

The parameters in the previous equations are defined in the following table:

ITEM	DESCRIPTION	SOURCE/VALUE
R	Range to surface	output of the re-tracker scheme and expressed in meter
Tx_Pwr	Transmitted Peak Power	Field 24 in SAR FBR format structure (and expressed in watts)
L_{atm}	Two Ways Atmosphere Losses	to be modelled and expressed dimensionless in linear scale
L_{RF}	Two Ways Residual Losses	to be characterized and expressed dimensionless in linear scale
l_0	Radar Wavelength	to be extracted from IPF database, default value 0.022084 m
c_0	Speed light in vacuum	299792458 m/sec
wf	footprint widening factor	1: no weighting window; 1.486 · rv in case of Hamming window
$R_\oplus$	Mean Earth Radius	6378137 m
G_0	Antenna Gain at Boresight	to be extracted from IPF database, default value 10^(4.26)
PTR_width	3db Range Point Target	to be extracted from IPF database, default value 2.819e-09 sec
V_s	Satellite Along Track Velocity	From Field 11 in SAR FBR format structure and expressed in m/sec
t_B	Burst Length	to be extracted from IPF database, default value 0.00352 sec
bias_sigma_0	CryoSat-2 Bias for s0 (db)	to be defined

4.3. Waveform Power Estimation

The waveform power, P_u, can be estimated by retracking techniques. Currently available retrackers for SAR Altimeters are mostly oriented for open-ocean and do not take into account the particularities of electromagnetic scattering from ground surfaces. In addition, the inhomogeneous nature of land surfaces prevents the use of scattering models, which could lead to erroneous outputs in the retracking process.

For these reasons, within the LOTUS project we have chosen the so-called "Offset Centre-of-Gravity" *(*OCOG*)* retracker, [Wingham et al., 1986], to estimate the total waveform power. The OCOG retracker is a simple and robust empirical retracker, which calculates the center of gravity (C) of the waveform energy. The different parameters related to the OCOG retracker are displayed in figure 3, where W and A represents the width and height of the rectangle centered around C with an area equal to the summed power in the waveform. L is the position of the leading edge.

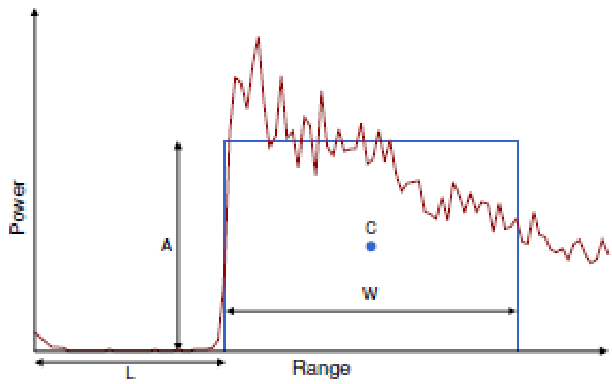

Figure 4.1: The OCOG retracker (adapted from Stenseng (2011))

$$C = \frac{\sum_n n\, p_n^2}{\sum_n p_n^2} \tag{10}$$

$$A = \sqrt{\frac{\sum_n P_n^4}{\sum_n p_n^2}} \tag{11}$$

$$W = \frac{(\sum_n P_n^2)^2}{\sum_n P_n^4} \tag{12}$$

$$L = C - \frac{1}{2} W. \tag{13}$$

Here *n* is the bin number and *P* is the power.

The obvious advantage of the OCOG retracker is that it works everywhere over all surfaces as it does not require a waveform leading edge as all other empirical and physical retrackers. Consequently, it will always enable the estimation of range to the surface and total received power by the radar. However, the disadvantage is that this retracker is very crude and very inaccurate. This retracker is frequently used when all other retrackers fail, and therefore it is significant for snow depth estimation, as the topography and inhomogeneity of the terrain could lead to non-conventional waveform shapes.

5. References

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