Estimating Terrestrial Snow Depth With the Topex–Poseidon Altimeter and Radiometer

Fabrice Papa, Benoît Legresy, Nelly M. Mognard, Edward G. Josberger, and Frédérique Remy

***Abstract*—Active and passive microwave measurements obtained by the dual-frequency Topex–Poseidon radar altimeter from the Northern Great Plains of the United States are used to develop a snow pack radar backscatter model. The model results are compared with daily time series of surface snow observations made by the U.S. National Weather Service. The model results show that Ku-band provides more accurate snow depth determinations than does C-band. Comparing the snow depth determinations derived from the Topex–Poseidon nadir-looking passive microwave radiometers with the oblique-looking Satellite Sensor Microwave Imager (SSM/I) passive microwave observations and surface observations shows that both instruments accurately portray the temporal characteristics of the snow depth time series. While both retrievals consistently underestimate the actual snow depths, the Topex–Poseidon results are more accurate.**

***Index Terms*—Microwave radiometry, Northern Great Plains of the United States, radar altimetry, snow depth.**

I. Introduction

SEASONAL snow cover (excluding snow over ice sheets and sea ice) has the largest areal extent and variability of any component of the cryosphere on the earth's surface. Most of the earth's snow-covered area is located in the Northern Hemisphere, and temporal variability is dominated by the seasonal cycle. Northern Hemisphere mean snow-cover extent ranges from approximately 3.8 million km^2 in August to 46.5 million km^2 in January [1]. Accurate large-scale measurements of snow extent, depth, and length of season are important hydrological and climatic parameters that are necessary to determine river discharge especially in the Arctic Ocean where this fresh water input greatly influences the ocean circulation and sea ice cover [2].

Snow pack thickness and extent and the duration of the snow period are important parameters to characterize and understand climate changes. Global average surface temperature has increased by approximately 0.5 °C since the last half of the ninet land in high latitudes [3].

In that context, changes in snow cover mean characteristics take a special significance. The impact of a warming affects the areal extent of continental snow cover and the length of the snow season. The sensitivity of snow cover extent and duration to temperature and its feedback effects [4], [5] suggest that snow cover would generally have a strong negative feedback relationship to global warming. For example, the apparent unprecedented global warming of the 1980s was accompanied by a retreat of the mean annual North American snow cover by 10% [6]. Using satellite observations of the Northern Hemisphere snow cover extent made since the late 1960s, Robinson *et al.* [7] documented this inverse relationship between hemispheric snow extent and surface air temperature, and Groisman *et al.* [8] demonstrated that snow exerted the strongest positive feedback on the radiative balance of the earth during the spring period. Goodison and Walker [9] presented other potentially important indicators of changes in snow conditions that reflect changes in climate. These include the changes in date of the beginning or end of continuous snow cover, in the number of days of continuous snow cover, in snow depth evolutions, or in spatial distribution of snow cover over Canada. These indicators suggest that spring snow extent should be a significant and sensitive indicator of hemispheric temperature changes. Recent efforts to expand the Northern Hemisphere satellite snow extent record with *in situ* data [10] provided evidence of a significant decrease in spring snow extent over Eurasia since 1915.

Satellite observations provide a unique means of monitoring snow cover characteristics and its variations on a regional or global scale. Several algorithms have been developed to estimate snow water equivalent or snow depths [11], [12], using combinations of brightness temperature measurements from the Scanning Microwave Multi-frequency Radiometer onboard Nimbus7 and the Satellite Sensor Microwave Imager (SSM/I) onboard the Defense Mapping Satellite Program satellite series [13]. Seasonal biases mostly due to the transparency of thin snow cover ($<$ 5 cm) in the microwave frequencies in the beginning of the snow season and the occurrence of melting events in the late season were clearly identified. These algorithms, developed to estimate snow depth, assume a spatially constant grain size throughout the winter.

Josberger and Mognard [14] derived an algorithm that includes a proxy for snow grain size growth. This algorithm, which was compared with National Weather Stations (NWS) *in situ* snow depth measurements in the Northern Great Plains (NGP) of the United States during the winter season 1996–1997, yielded better estimates than the other algorithms that assume constant grain size in time and space. For all passive microwave snow pack algorithms, the date for the beginning and the end of the seasonal snow pack contains significant errors, and problems of over- and underestimation of snow depth still occur. A recent study [15] developed a

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technique to determine the presence of snow using European Remote Sensing (ERS) satellite radar altimeter measurements from the NGP. The technique is based on a reduction of "noise" in the returned signal. Furthermore, a comparison between radar altimeter measurements in Ku-band and snow depth from NWS demonstrated the potential of such methods to recover qualitative estimations of snow pack extent and depth based on the attenuation of the radar signal. This paper investigates the opportunities offered by the dual-frequency Topex–Poseidon radar altimeters to improve the measurement of snow pack properties from satellite sensors. The datasets used in this analysis are presented in Section II, and the methods to estimate snow parameters are explained in Section III. Section IV presents the results obtained over the NGP, and Section V presents a discussion of the results.

II. Datasets

The Topex–Poseidon altimeter is a dual-frequency instrument (Ku-band, 2.3-cm wavelength, and C-band, 6-cm wavelength with a footprint of a few kilometers) with a 10-day repeat cycle (instead of 35 days for ERS altimeter). Details about the Topex–Poseidon altimeters can be found in [16] or [17]. Even for continental targets, where variations in the surface elevation are important, measurements with both frequencies can be considered simultaneously over a time interval of 0.1 s, corresponding to 700-m along-track distance [18]. A nadir-pointing microwave radiometer at 18, 23, and 37 GHz (with footprint of 42, 35, and 22 km, respectively) operates simultaneously on the same platform to initially provide tropospheric corrections to the altimetric height measurements. The presence of the two nadir pointing instruments onboard the same platform offers the opportunity to compare simultaneous active and passive microwave measurements.

Four types of measurements are used in this study: the backscattering coefficients in C- and Ku-band from the Topex–Poseidon (T–P) radar altimeters; the 19- and 37-GHz horizontal polarization brightness temperatures measurements from the SSM/I; the 18- and 37-GHz brightness temperatures measurements from the radiometer onboard Topex–Poseidon; and the daily *in situ* measurements of snow depths and temperatures from five NWS stations located less than 50 km away from the altimeters footprint. The active and passive microwave observations from T–P instruments were averaged every 0.2° in latitude and longitude and compared with the five NWS sites located close to the satellite tracks.

The study focuses on the 1996–1997 snow season, when record snowfalls occurred over the NGP and when the area along the Red River in North Dakota had maximum snow depths of 1 m or more. The NGP are well suited for this study because it consists primarily of open farm land or prairie, has little topographic variation, is subject to very cold temperatures, and has more than 280 stations that report the snow depth on a daily basis. This area has been used for numerous passive microwave snow pack studies, where the extensive *in situ* observations have been interpolated to the passive microwave grid (25×25 km^2) using the geostatistical process called krigging.

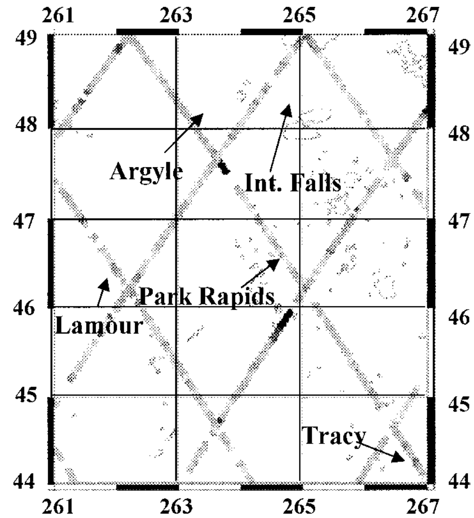

Fig. 1. Map of the NGP with the T–P satellite tracks and the localization of the five NWS sites.

Fig. 1 shows the portion of the NGP used in this study along with the satellite tracks and the location of the five NWS sites

III. Methods

In two recent papers [15], [19], we analyzed four ERS2 radar altimeter parameters derived from the altimeter waveform [20]: the altimetric height, the backscattering coefficient, the width of the leading edge, and the slope of the trailing edge. For the NGP, during the extreme winter of 1997, the backscattering coefficient was found to be the most sensitive to the presence of snow. First, the snow indeed decreased the backscattering coefficient high-frequency variability along the satellite track due to the smoothing effect of snow cover [15]. Second, the mean value of the backscattering coefficient decreased as the snow cover thickened through the winter. The extinction coefficient, calculated from the backscattering coefficient and the NWS-measured snow depths, is close to the theoretical values (~ 1 m^{-1}) given in [21].

A. Modeled Backscatter for Snow

The averaged backscattering coefficient over the altimeter radar footprint is a sum of three different effects: reflection from the snow surface, the volume-scattering effect of the entire snow layer, and two-way attenuation of the ground return signal. With the ERS2 single -frequency radar altimeter, one cannot differentiate the ground reflection attenuation by the snow from the air–snow interface echo contribution. The dual-frequency T–P instrument offers the opportunity to differentiate and analyze these two effects.

Previous studies over Antarctica (the only region where an extinction coefficient has been calculated) showed that volume scattering may contribute to half of the total backscattering coefficient if the snow layer is several meters deep [22]. This is not the case for the NGP where snow depths never exceeded 1.5 m, and where the volume-scattering contribution can be neglected when compared with the two other terms.

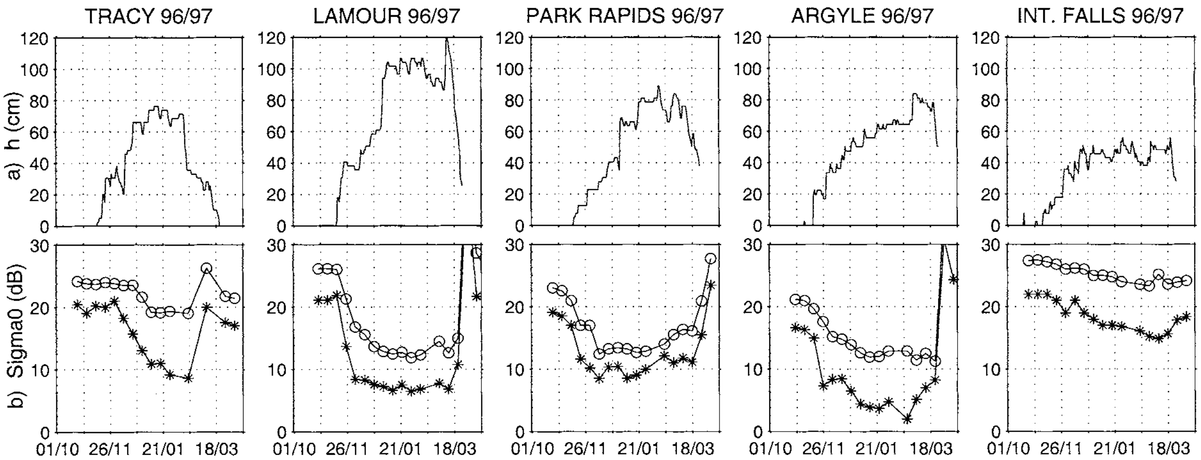

Fig. 2. (a) Snow evolution from NWS *in situ* data in centimeters. (b) Backscattering coefficients evolution in C-band (o) and Ku-band (*) in decibels.

In the absence of snow cover, the mean backscattering coefficient σref is the ground signal

$$\sigma\text{ref} = 10 \ln (\sigma\text{ground}). \tag{1}$$

In the presence of snow, the mean backscattering coefficient σtot can be approximated by the sum of the surface echo and the ground signal attenuated by the presence of the snow as follows:

$$\sigma\text{tot} = 10 \ln (\sigma\text{ground} * \exp(-2\, ke\, h) + \sigma\text{surface}) \tag{2}$$

where ke is the extinction coefficient in 1/meters; h is the snow depth in meters; and $\sigma\text{surface} = f(F^2/S^2)$, where F is the Fresnel coefficient, and S is the roughness of the surface.

Combining (1) and (2) gives the backscattering coefficient difference between the snow free and the snow covered area

$$\frac{1}{(2\chi)}(\sigma\text{ref} - \sigma\text{tot}) = ke * h - \frac{1}{2} \ln \left(1 + \frac{\sigma\text{surface}}{(\sigma\text{ground} * \exp(-2\, ke\, h))}\right) \tag{3}$$

with $\chi = 4.34$, due to the transformation of the neperian logarithm into a decimal logarithm. The temporal evolution of the backscatter with respect to σref gives information on the snow pack characteristics during the winter season.

B. Topex–Poseidon Dual-Frequency Radar Altimeter Over the NGP

The temporal evolutions of the backscattering coefficients in C- and Ku-bands are presented in Fig. 2. Fig. 2(a) displays the snow depth evolution measured at the NWS stations, and Fig. 2(b) gives the collocated backscattering coefficient in C– and Ku-bands.

At the start of the snow season (end of October, beginning of November), the values of the backscattering coefficients in C- and Ku-bands begin to decrease. As observed in a previous study [19], the evolution of the backscattering coefficient is a good parameter to detect and map the onset of snow cover.

For the five sites, the backscattering coefficient decreases strongly for both frequencies as the snow depth increases. Also, note that the backscattering in C-band is less sensitive to shallow dry snow than the Ku-band (the shallow dry snow being transparent at lower frequency), and second that the Ku-band is noisier than the C-band. Moreover, the dual-frequency altimeter shows trends that differ as a function of frequency for each station.

TABLE I
RESULTS OF THE MODELED BACKSCATTER FOR SNOW COVER

	C band		Ku band	
	σ_surface (dB)	*ke* (m-1)	σ_surface (dB)	*ke* (m-1)
Tracy	18	1.6	7	2.7
Lamour	12	2.4	7.6	3.4
Park Rapids	12,2	2.4	9.5	3.2
Argyle	2	1.6	2	2.5
Int. Falls	1	0.8	1	1.4

For Lamour and Park Rapids the signal saturates, especially in Ku-band where the backscattering coefficient decreases from 21–8 dB as the snow depth increases to 50 cm, but then levels out as the snow depth continues to increase to more than 1 m.

On the contrary, the backscattering coefficient from Tracy, Argyle, and International Falls does not saturate, but shows the following behavior.

- *Tracy:* Ku-band backscattering coefficient decreases from 20 dB to around 10 dB, and the C-band decreases from 25–19 dB as the snow depth increases to 75 cm.
- *Argyle:* In Ku-band, the signal decreases from 17–5 dB and, in response to the 80-cm thickening snow cover, from 21–12 dB in C-band.
- *International Falls:* Backscattering coefficient decreases from 22 to about 16 dB for Ku-band and from 27–22 dB for C-band as the snow depth increases to 50 cm.

For these three last stations, deeper snow corresponds to a larger decrease in the backscatter coefficient, and the C-band is still less attenuated than the Ku-band for same snow pack characteristics, since low frequencies penetrate more than higher frequencies in the snow [21].

C. Validation of Modeled Backscatter for Snow Cover With T–P Measurements

Table I shows the results of (3) used to compute the extinction coefficient for both bands using the time series of snow

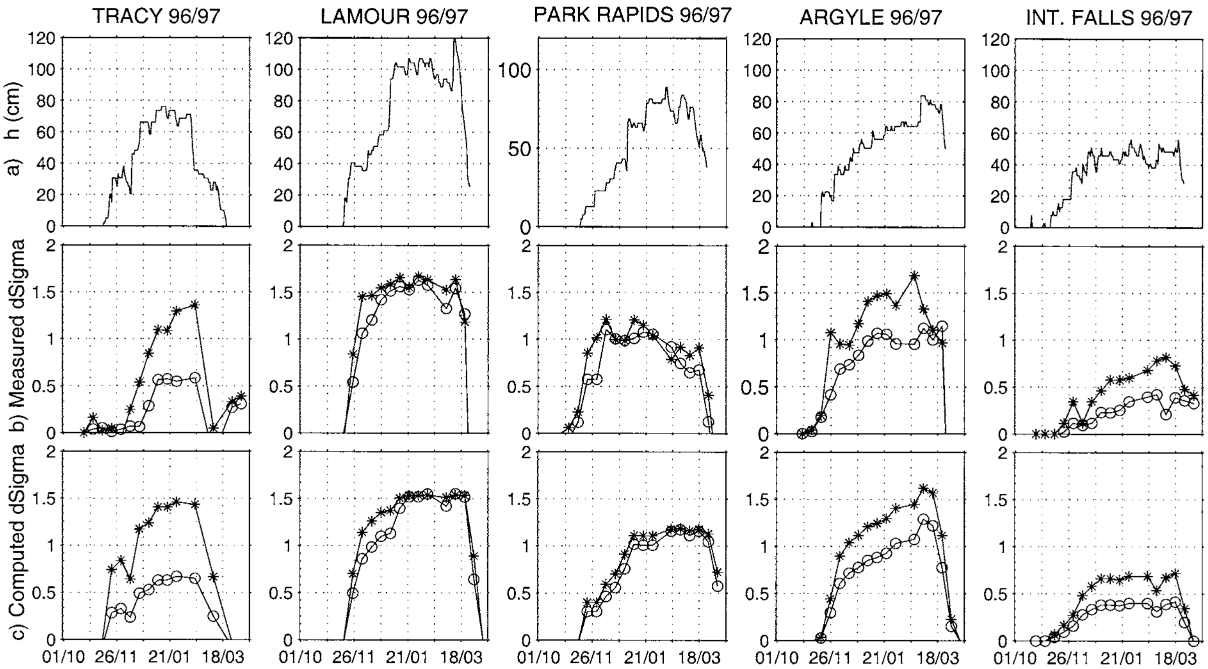

Fig. 3. (a) Snow evolution from NWS *in situ* data (centimeters); (b) dSIGMA $= 1/(2\tau) * (\sigma\_\mathrm{ref} - \sigma\_\mathrm{tot})$ evolution from the measured altimetric data; and (c) dSIGMA $= 1/(2\tau) * (\sigma\_\mathrm{ref} - \sigma\_\mathrm{tot})$ evolution from the modeled backscatter.

depth from each of the five sites. The calculated values of the extinction coefficient are in the range of the theoretical value given by [23] for both frequencies. For the five stations, Fig. 3(a) shows the time series of the *in situ* snow depth; Fig. 3(b) shows the measured backscattering coefficient difference dSIGMA $= 1/(2\chi) * (\sigma\mathrm{ref} - \sigma\mathrm{tot})$; and Fig. 3(c) shows the modeled backscattering coefficient difference dSIGMA using the computed values of ke and the time series of snow depth from each NWS station.

The temporal behavior of the modeled backscattering coefficient is in good agreement with the observed backscattering coefficient during the snow season. The saturation phenomena at Lamour and Park Rapids are well reproduced, and for the three other stations, the modeled backscattering coefficient agrees with the measured backscattering coefficient.

The modeled backscattering coefficient enables us to understand the physical processes that lead to the altimetric response to snow cover.

IV. Results

A. Snow Depth Estimations With Topex–Poseidon Radar Altimeter

To estimate snow depth from the altimeter measurements, we use (3) where the term $ke * h$ is the only unknown. Since the backscattering coefficient decreases when the snow cover starts, we estimate from the altimetric measurements the following terms:

- $1/(2\chi) * (\sigma\mathrm{ref} - \sigma\mathrm{tot})$;
- σground;
- σsurface;

and we obtain for each date and each station a value of $ke * h$. To reduce the altimetric noise level, we smoothed the backscattering coefficient in each band with a median filter. The ground backscattering coefficient σref is estimated with an accuracy of approximately 1 dB, which corresponds to the root mean-square high-frequency value in the absence of snow. In case of saturation of the backscattering coefficient (for Lamour and Park Rapids stations), σsurface is estimated for both frequencies as the minimum value reached during the snow season.

Since we do not have information on the value of ke, we have assumed that, for each band, ke is constant and has the same value for each test site. Using the work in [21], we chose a mean value of 1.8 m^{-1} for C-band and 2.5 m^{-1} for Ku-band. These assumed values are in the range of the extinction coefficients determined in the previous modeling section. The *in situ* and altimeter-derived snow depths are plotted for the five test sites (see Fig. 4).

Comparing the derived snow depth time series with the observed time series shows the following. First, the beginning and the end of the winter snow pack are accurately determined within the repeat cycle of the satellite. The altimeter measures snow during the first pass when snow is present on the ground except for Tracy, where the altimeter misses the beginning on the first pass when snow was present, detecting snow only on the second pass ten days later. The end of the snow pack is also well reproduced except for Tracy, where the altimeter estimated that snow had disappeared at the end of February when there was still snow on the ground, but where again measuring snow at the end of the snow season after March 20. This could be the result of warming that produced liquid water in the snow pack that would greatly alter the electromagnetic properties of the snow pack.

Next, while the snow depth retrievals from both bands exhibit a temporal behavior quite similar to the NWS measurements, the Ku-band snow depth retrievals are generally more accurate than the snow depths computed using C-band. Table II

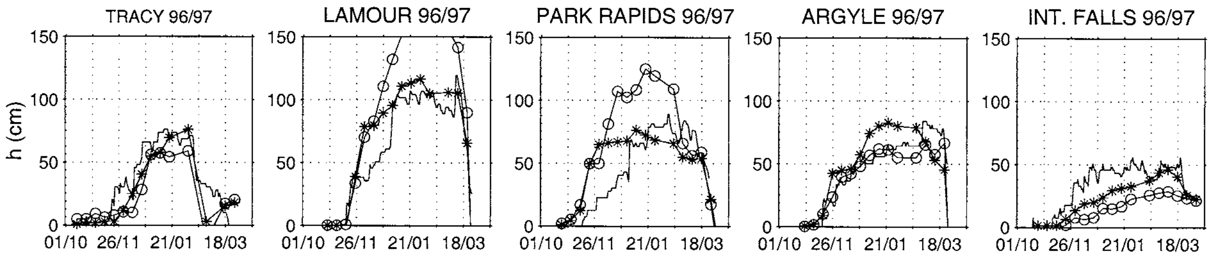

Fig. 4. Snow depth h (centimeters) estimated with the modeled backscatter inversion for the C-band (o) and Ku-band (*) versus the snow depth from *in situ* data (centimeters).

shows the mean difference between the Ku-band altimeter-derived snow depth estimation and the *in situ* measurements. For Lamour and Park Rapids (the two stations where the backscattering coefficients saturate), the C-band-estimated snow depths are over estimated by about 50 cm compared with the *in situ* measurements and to the Ku-band estimates. For the three other stations, C- and Ku-band-derived estimates are close to the *in situ* measurements with C underestimating the snow depth by up to 40 cm for International Falls compared with Ku.

Compared with the Ku-band, the C-band-retrieved snow depth estimations are in poor agreement with the *in situ* measurements for Lamour and Park Rapids. This can be explained by the lack of information on snow pack parameters, which prevents us from evaluating the extinction coefficients ke for each station. More accuracy in the snow depth estimation would result from information on the snow grain size and the snow density at each site.

B. Comparisons of Snow Depth Algorithms Using T–P and SSM/I Radiometers Measurements

This section compares the observations from the nadir-pointing radiometer onboard Topex–Poseidon with those made by the SSM/I radiometer that operates at an angle of 52° of incidence. Several algorithms are currently available to evaluate snow cover and snow depth parameters for specific regions and specific seasonal conditions. The snow cover and snow depth study over the NGP presented here were generated using the algorithm developed in [11]. The difference in brightness temperature between the SSM/I 37- and 19-GHz horizontally polarized channels is used to derive snow depth over the five NGP stations according to the relation

$$h = 1.59(T19_H - T37_H) \tag{4}$$

where h is the snow depth in centimeters, and $T19_H$ and $T37_H$ are the horizontally polarized brightness temperatures in degrees Kelvin.

Since the T–P radiometer senses at the nadir, and the SSM/I has a 52° incidence angle, we corrected for the 52° incidence angle difference and obtained the following relationship for the Topex–Poseidon data:

$$h = 1.59\frac{(T19 - T37)}{\cos(52°)}. \tag{5}$$

Fig. 5 shows both the *in situ* snow depth measurements for the 1996–1997 snow season and the estimates derived from the SSM/I and T–P radiometers. The results of the comparison are

TABLE II
RESULT OF THE MEAN DIFFERENCE BETWEEN THE SNOW DEPTH ESTIMATED FROM TOPEX-POSEIDON AND THE *IN SITU* DATA AND THE STANDARD DEVIATION

	Mean(m)	Standard deviation (m)
Tracy	-0.063	0.12
Lamour	0.038	0.35
Park Rapids	0.030	0.30
Argyle	0.018	0.24
Int. Falls	-0.074	0.11

shown in Table III. The snow depths derived from both sets of radiometric data consistently underestimate the snow depth, but both estimates correlate well with the measured snow depths $0.69 < R < 0.94$. For Tracy and International Falls, SSM/I poorly estimates the snow depth evolution ($b1 = 0.23$, $R = 0.69$, $b1 = 0.45$, and $R = 0.76$ respectively), while the T–P radiometer is better with $b1 = 0.95$ and $R = 0.85$ for Tracy and $b1 = 0.87$ and $R = 0.74$ for International Falls. For Lamour and Park Rapids, both radiometers exhibit a similar temporal behavior, and they greatly underestimate the snow depth. These two stations are the ones that exhibited saturated radar signals.

V. DISCUSSION

A comparison of snow depths determined from the T–P radiometers, the SSM/I radiometers, and the *in situ* measurements shows that the radiometer onboard Topex–Poseidon yields more accurate snow depths. The differences may be the result of the different incidence angles between the two instruments, which might infer shadowing phenomena, due to vegetation or topography as well as to the difference in footprint size.

When comparing active and passive microwave responses, two different trends clearly appear.

- When the radar signal saturates (for Lamour and Park Rapids), the snow depth determinations from both radiometers are equivalent, and both underestimate the snow depths.
- When the backscattering coefficients in C- and Ku-bands are different and do not saturate (for Tracy, Argyle, and International Falls), the SSM/I-derived snow depths compare poorly to snow depths derived from the Topex–Poseidon radiometer.

This behavior may be the result of the different contributions of the surface signal. For the radar, the surface return is directly related to the relationship between the Fresnel coefficient F and the roughness of the surface S. Different scenarios for F and S should be verified during ground campaigns.

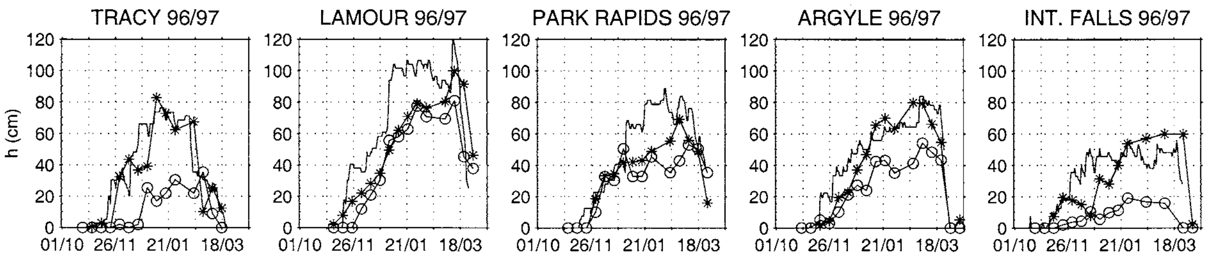

Fig. 5. Snow depth estimation h (centimeters) from Topex–Poseidon radiometer (*) and SSM/I radiometer (o) versus *in situ* data (centimeters).

TABLE III
REGRESSION AND CORRELATION COEFFICIENT FOR Y VERSUS X; $Y = b0 + b1 * X$; R CORRELATION COEFFICIENT. (a) SSM/I RADIOMETER VERSUS *IN SITU* DATA; (b) T–P RADIOMETER VERSUS *IN SITU* DATA; (c) SSM/I RADIOMETER VERSUS T–P RADIOMETER

SSM /I radiometer versus in-situ data (a)

	b0	b1	R	Error b0*0.1	Error b1*10-2
Tracy	-0.44	0.33	0.69	4.09	9.6
Lamour	-7.7	0.73	0.86	5.31	6.9
Park Rapids	4.80	0.57	0.82	5.53	10
Argyle	-0.08	0.71	0.93	5.1	9.7
Int. Falls	-0.07	0.45	0.76	0.5	10

T-P radiometer versus in-situ data (b)

	b0	b1	R	Error b0*0.1	Error b1*10-2
Tracy	-0.51	0.95	0.85	6.92	16
Lamour	2.41	0.75	0.94	8.9	11
Park Rapids	-4.47	0.83	0.93	4.51	8.8
Argyle	0.13	0.98	0.94	6.1	1.6
Int. Falls	-2.25	0.87	0.74	1.3	3

SSM/I radiometer versus T-P radiometer (c)

	b0	b1	R	Error b0*0.1	Error b1*10-2
Tracy	3.7	0.34	0.53	4.37	10
Lamour	-2.32	0.96	0.93	5.69	9.55
Park Rapids	7.80	0.64	0.88	4.07	10
Argyle	2.2	0.59	0.95	3.38	6.4
Int. Falls	4.67	0.42	0.93	2.09	5.18

1) If F is small (dry snow conditions), the surface roughness variation is not a strong contributor, and the surface return is small when compared with the attenuation through the snow layer.
2) If F is not small (wet snow conditions), two cases can occur.
   a) *The roughness of the air–snow interface S has a strong value:* The surface echo term is small and can be neglected when compared with the attenuation by the backscattering coefficient.
   b) *The surface is not rough: S* has a small value; the surface echo is the dominant term; and it is the source of the saturation phenomena.

The behavior of the backscattering coefficient for both frequencies represents, thus, a good parameter to estimate the contribution of the surface effects, which also play an important role in the passive microwave measurements over snow-covered areas.

Ulaby *et al.* [23] show the brightness temperature dependence on the angle of incidence, the roughness and wetness parameters. For the 10.7-GHz frequency (frequency used by Ulaby *et al.*, but not available for the satellite radiometers considered here), the roughness of the snow has strong effects on the brightness temperature for angles of incidence above 60°, whereas for the 37-GHz frequency, the surface effects become significant at angles greater than 30°. For both rough and smooth snow surface conditions, the 37-GHz frequency) at nadir has the same brightness temperature at about 263 K. For an angle of incidence of 51°, the brightness temperature for a smooth snow surface is 242 K, while it is 257 K for a rough surface. This 15-K difference, for the same snow depth and grain size, may explain the differences in the snow depth estimations from the two passive microwave sensors.

VI. CONCLUSION

This paper investigated the response of active and passive microwave measurements to snow cover and its evolution over the winter of 1996–1997 for five stations located in the Northern Great Plains of the United States. The comparison and subsequent merging of data from these two types of measurements is a means to improve our understanding of microwave scattering by snow packs that will lead to more accurate estimates of snow pack thickness from satellite observations. First, the inversion of a model of snow cover backscatter for the dual-frequency radar altimeter measurements reasonably reproduces the onset, the growth, and the end of the snow pack through the winter.

Second, a comparison between passive microwaves measurements from Topex–Poseidon and the SSM/I satellite to estimate snow depth shows that a radiometer at a nadir-pointing angle yields more accurate estimations than a radiometer at 52° incidence angle. Josberger and Mognard [24], [25] show that the spatial variation in snow pack grain size can result in erroneous snow depths when calculated from conventional passive microwave algorithms. Future more accurate snow pack property algorithms must include grain size variations, both temporally and spatially. The combination of SSM/I passive microwave observations with multifrequency radar observations can be the basis for such an algorithm. SSMI observations provide large-scale coverage and a repeat observation time of one to three days that is necessary for hydrologic and climate studies, but the information in the single channel of data (19–37 GHz) is insufficient to unravel the complexities of natural snow packs. The T–P radar observations provide additional information at two other frequencies, but at a repeat cycle of only ten days. The additional snow pack information contained in the radar observations may then be used to provide an ongoing correction to passive microwave snow pack retrievals.

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