ANTARCTIC ICE SHEET SNOW PROPERTIES DERIVED FROM ERS ALTIMETER DATA.

Benoît LEGRESY and Frédérique REMY

UMR5566/GRGS (CNES-CNRS-UPS), 18 Av. E. Belin, 31401 Toulouse cedex 4, FRANCE

Tel : (33) 5 61 33 29 56 , Fax : (33) 5 61 25 32 05

E-mail : legresy pontos.cst.cnes.fr

ABSTRACT

The ERS-1 radar altimeter allows the observations of 80% of the Antarctica ice sheet. The radar observations (both waveform shape and backscattering coefficient) give useful information on snowpack characteristics, such as surface roughness, internal stratification or ice grain size. Snow surface conditions get spatio-temporal variations that strongly affect both the radar waveform, backscattering coefficient, and the retrieval of the height from the altimetric waveform. These temporal effects are to be taken into account to survey ice sheet mass balance. However, they bring very useful new indications on the snowpack characteristics that may thus be retrieved despite the large set of parameters involved in the radar echoes. An analysis of the temporal and geographical signal present in the data provides global maps of penetration depth.

INTRODUCTION

The launch of the ERS-1 satellite gave new impetus to satellite altimetry work over the polar ice-caps. It is now possible to accurately map 80% of Antarctica and virtually all of Greenland. ERS-1's two shifted geodetic cycles provide very high spatial resolution: inter-track spacing of 4 km at 60° latitude and 2 km at 76°.

Observing the polar ice caps via altimetry meets at least three distinct objectives:

- constraining ice flow models by studying the shape of the ice sheet, that is the subject dealt with in Remy & Legresy (this issue),

- monitoring their mass budgets by successive topographic surveys, requiring 10-cm accuracy,

- describing and monitoring trends in climatic surface characteristics (type of snow, surface roughness, and so on) by analyzing radar echoes.

On the polar ice caps the altimeter height can be biased by the penetration of radar waves into the snowpack. Estimating the bias is difficult because of the many factors affecting surface and volume backscattering. The radar echo depends on the surface micro-roughness, the kilometer-scale roughness, the stratification of the snowpack, the radar extinction in the snow, and the antenna pattern. The echo therefore depends on the frequency used. A way to better understand the altimetric signal is to increase the volume of altimetric parameters relative to the number of unknowns. This was done for instance using TOPEX dual-frequency altimeter that provides a means to detect the induced bias empirically (Rémy et al., 1996). However, TOPEX/POSEIDON covers only a small part of Greenland and does not cover Antarctica.

1. SPATIAL behavior of the altimetric signal.

Data from the 35-day cycle n°87 of ERS1 including about 500 tracks in ocean mode have been analyzed here. Waveforms are analyzed at the 20 Hz rate (i.e. each 350 m along track). Three principal parameters are extracted from the individual waveforms in addition to the height measurement using a fitting process (Legresy & Remy, 1997a). The total energy return is related to the backscattering coefficient s^0 (expressed in dB). The half leading edge width, named Tr (expressed in altimetric gates, e.g. 47 cm for ocean mode) corresponds to the first part of the impact of the radar wave on the ground surface, e.g. one kilometer scale. The trailing edge slope, further named Fl (expressed in 10^{-4} Neper/gate), This latter corresponds to the whole altimeter footprint scale.

Large scale maps of these parameters were constructed (Figure 1, 2 and 3) by averaging the parameter values over 50*50 km^2. The three maps show important and coherent signals. A principal component analysis carried on the hole ice sheet dataset indicates that one principal mechanism controls the altimetric response of the surface, explaining 89% of the variance of the three parameters. It indicates that when s^0 diminishes, Tr and Fl increase. Fl is controlled by the ratio between volume and surface signal if averaged over a 50 km scale while s^0 is controlled by the summation of both terms at the same scale. At this scale Fl is also dependent on the surface slope and curvature while s^0 is little affected by undulations and slope. In this way, the decrease in the ratio volume/surface associated with the increase of their sum can only be explained by an increase in surface backscattering. Note that a control of the total backscattering by the extinction coefficient as it is currently admitted (Davis and Zwally, 1993) would enhance s^0 and Fl together when Figure 1, 2 and 3 clearly suggest a control by surface backscattering.

Generally, in high accumulation zones, e.g. coastal zones of East Antarctica and the eastern part of West Antarctica, the s^0 is weak and Tr and Fl are high. Over these regions, the important surface microroughness caused by strong winds leads to low surface backscattering. Indeed, Remy et al. (1990) showed, with Seasat data, in the [90-150 °E ; 66-72 °S] region, that the wind influences the surface microroughness and concluded that the backscattering is inversely proportional to the wind intensity.

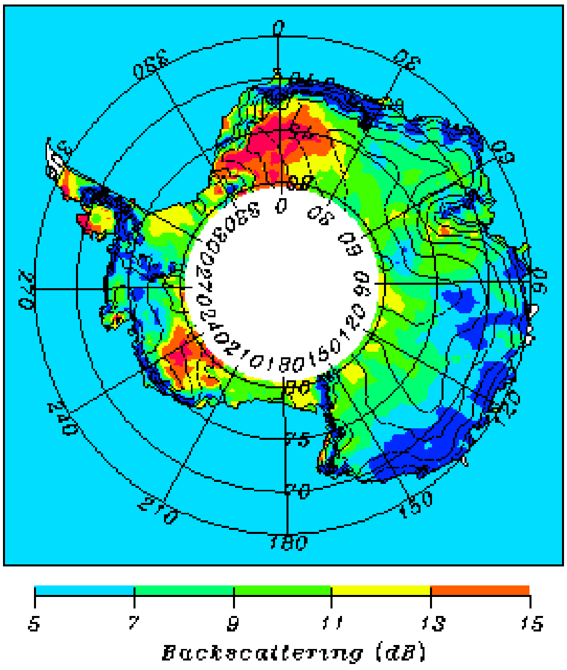

Figure 1 : Map of the backscattering coefficient (in dB). The backscattering varies from few dB in Adelie and Wilkes Land to more than 17 dB in the Queen Maud Land. This large range of values reveal the huge variety of responses of the ice cap surface to radar waves.

Variations in backscattering are then due to variations in surface microroughness representative of wind intensity. The leading edge is high, partly because of the important macro-roughness and partly because it is enhanced by volume echo. The trailing edge is particularly raised, partly by the surface slope (Legresy, 1995), and partly by the importance of volume echo compared with surface echo.

Elsewhere, for instance, at 120°E, from the coast to the interior, the accumulation varies from 50 cm/year (water equivalent) to 5 cm/year. The wind is particularly high along the coast (>18 m/s) while it is less than 6 m/s in the interior. Thus, the surface roughness (micro and macro) is very important toward the margin and it is low in high altitudes. The important surface signal compared to subsurface one and the decreasing surface slope make the trailing edge go down to its reference value Higher in latitude the backscattering becomes very high, the leading edge is short and the trailing edge slope becomes lower than its reference value. This can be explained by a particular surface echo process that could be described as intermediate between rough surface echoes observed above ice caps and specular surface echoes observed above sea ice. It is not possible here to conclude if it is caused by a particular distribution function of surface roughness slope, or by particular effects of near surface stratification in this low accumulation zone.

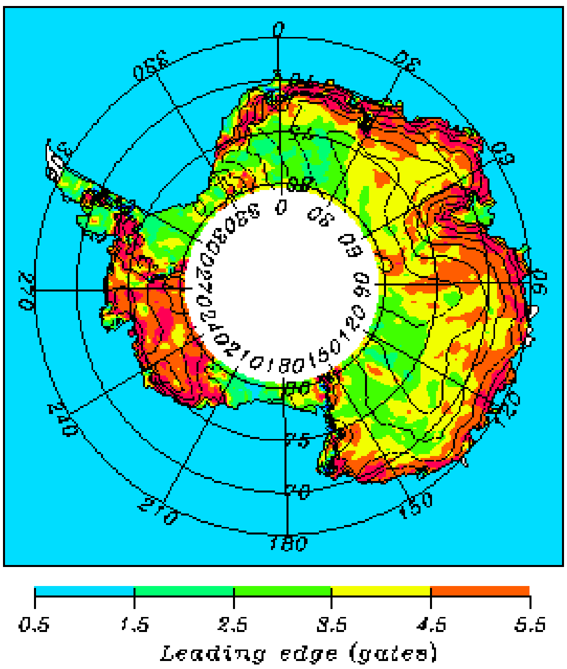

Figure 2 : Map of the Leading Edge Width of the altimetric waveform (in altimetric gates, i.e. 47 cm in ocean mode used here). It varies from few gates (0.5 being the minimum value) in the interior of the continent where there is quite no surface macro-roughness like snow dunes to more than 6 gates near the coast where the strong winds create surface patterns like sastruggi.

In the latter region, Rott et al. (1994) reported high backscattering coefficient with ERS1 scatterometer data in C band (5.2 GHz) at 35° incidence angle. They explained this high backscattering by important stratification. However the penetration depth (related to frequency) is more important for the scatterometer, while surface backscattering sensitivity (related to incidence angle) is more important for the altimeter. The scatterometer signal is thus largely controlled by subsurface characteristics.

On the contrary, in the Queen Maud Land (around 0° longitude) ERS1 scatterometer data (Rott et al., 1994) show low backscattering while ERS1 altimeter data show very high s^0 and low Fl. Indeed, such altimetric behavior can only be explained by a surface crust that can be transparent for scatterometer measurements.

In the West Antarctica Ice Cap, near 230°E, there is another region of high backscattering and low trailing edge, but in this case, the leading edge remains high.

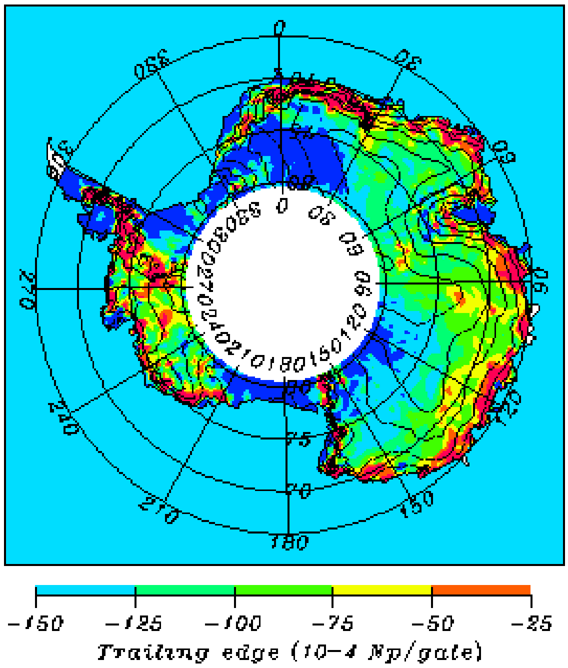

Figure 3 : Map of the Trailing Edge logarithmic Slope of the altimetric waveform (in 10^{-4} Np/gate, the nominal value over flat surfaces like ocean being around -130). It is raised by penetration and surface slope near the coast and it is particularly low in the central parts of the continent (e.g. in the Queen Maud Land) indicating, combined to the high backscattering, intermediate echoes between rough and specular surfaces.

2. USING THE ALTIMETRIC VARIABILITY TO CHARACTERIZE THE SNOWPACK CHARACTERISTICS.

We used the numerical echo model already developed in Legresy & Remy (1997a). This model allows to simulate waveforms from a fine digitized topography in order to have as realistic waveforms as possible, taking the satellite characteristics (altitude, antenna pattern,...) and the surface backscattering (s_s) into account. Volume echo is included using a simple model controlled by the contribution of a "one gate equivalent layer" to backscattering (s_v), resulting from stratification and ice grains scattering, and by the extinction through the snowpack (x_e) or equivalently the penetration depth (d_p), resulting from absorption and ice grains scattering. The outputs of this model are waveforms, which are retracked as ERS1 ones, all the waveform parameters are then recovered : s^0, leading edge width and trailing edge slope. Regarding to the temporal variability observation, it appears that the temporal signal at short time scale is linked with variations in s_s and that pertinent parameters describing this phenomenon are dTr/ds^0 and dFl/ds^0 (Legrésy & Rémy, 1997b). We thus modelize the behavior of these ratio with respect to variations in s_s, keeping the volume echo constant.

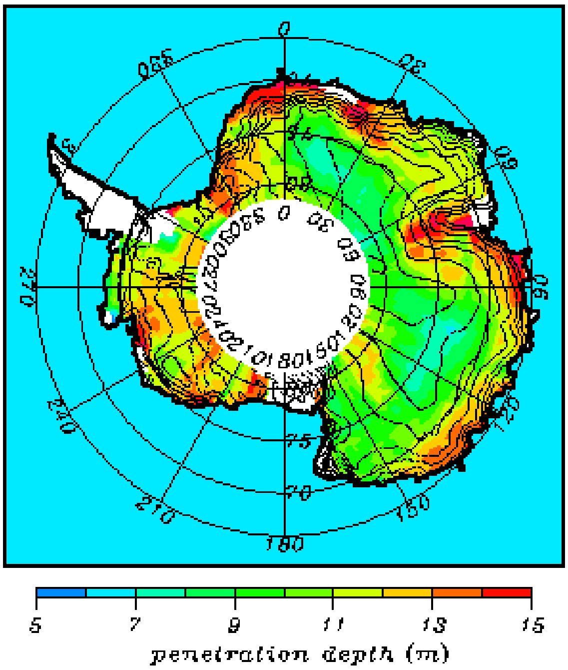

Figure 4 : Map of the Penetration Depth of the radar waves inside the snowpack obtained by inversion of the variability of the waveform parameters and backscattering through an echo model. It reveal low penetration (around 6m) in high altitude regions reaching more than 12 m in other regions.

These relations Tr(s^0) and Fl(s^0) are then processed for real topographic patterns and for couples (s_v/s_{ref} ; d_p) ranging *as* (0-0.5 ; 0-20 m). Both trailing edge and leading edge decreases while surface scattering increases. Surface scattering fluctuations greatly act on the trailing edge slope when the volume part is important, but have less effect when the volume part is weak. On the opposite, surface scattering fluctuations mostly act on the leading edge width. The fluctuations of both leading edge and trailing edge due to change in surface backscattering get a different behavior with respect to volume signal characteristics. This allows us to recover both volume scattering and penetration depth.

A linear inverse process using least square minimizing thus produces maps of penetration depth (Figure 4). The penetration depth values vary from 7 m inside the continent to 13 m at lower altitude. These values are in good accordance with previous ones. Ulaby & al. (1986) list penetration depths into dry snow, with 0.24 density and 0.5 mm ice grain size, of about 10m at 13 Ghz, while Ridley and Partington (1988) and Davis & Zwally (1993) with the help of Seasat altimeter waveform analysis respectively found penetration depth of 8m or varying from 5 to 10 m in small area, north to 72°S in Wilkes Land.

The extinction coefficient is due to absorption and scattering. Absorption is mostly controlled by the temperature and decreases from the coast to the dome. Scattering is mostly controlled by grain size increasing from the coast to the dome. The great scale behavior of figure 4 thus suggests that extinction coefficient geographical variations are mostly controlled by scattering.

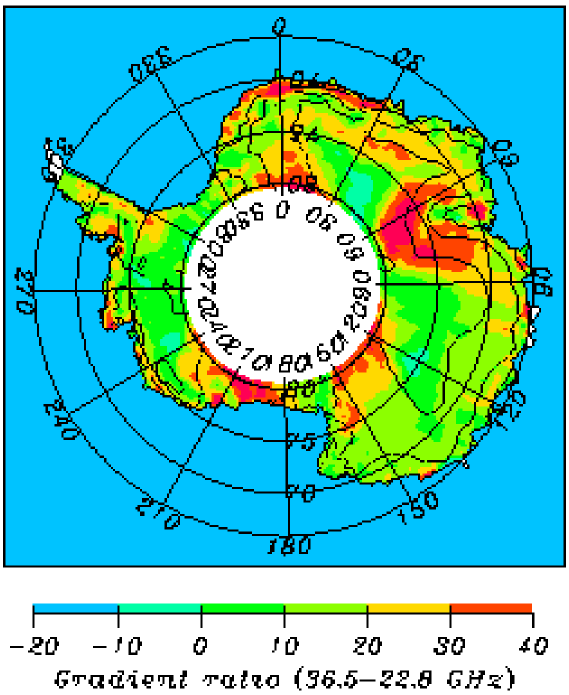

Figure 5 : Map of the Gradient ratio (in 10^{-3}) corresponding to the ratio between the difference and the sum of Brightness temperatures at 36.5 and 22.8 GHz measured by the ERS1 MWR. It varies from -20 in high altitude regions to more than 40 in the convergence zone near 60°E.

This latter assertion is confirmed by the gradient ratio map (Figure 5) obtained within the ERS1 radiometer data. This ratio is found to be correlated to the grain size (Surdyk & Fily, 1993) : the lower the ratio, the greater the grain size. The greater grain size increases the scattering and lower the penetration depth. This map is very coherent with the penetration depth in East Antarctica. This is however not verified in West Antarctica where the higher temperature perturbs the gradient ratio significance (Surdyk & Fily, 1993).

CONCLUSION

This study shows that it is possible to characterize the snowpack over Antarctica using radar altimetry. These observations are coherent with those made by the ERS scatterometer. It is possible to detect the penetration depth of radar waves into the snowpack using radar altimetry. The measured penetration depth are coherent with the dependence of the brightness temperature measured by ERS in the frequency. The launch of the Envisat satellite with a dual-frequency radar altimeter (Ku and S), offering the same coverage as ERS, brings the promise of great progress in altimetry over the ice caps.

REFERENCES

Davis C.H., H.J. Zwally 1993. Geographic and seasonal variations in the surface properties of the ice sheets by satellite-radar altimetry. J. of Glaciol., 39, 133, 687-697.

Legrésy, B. 1995. ‘Etude du retracking des formes d’onde altimétriques au-dessus des calottes polaires’. CNES report for ESA CT/ED/TU/UD/96.188 contract n° 856/2/95/CNES/006.

Legrésy B. and F. Rémy, 1997a, Surface characteristics of the antarctic ice sheet and altimetric observations, J. of Glaciology, in press.

Legrésy B. and F. Rémy, 1997b, Using the variability of the radar altimetric signal to map surface characteristics of the antarctic ice sheet , submitted to J. of Glaciology.

Rémy, F., C. Brossier and J.F. Minster, 1990, Intensity of a radar altimeter over continental ice sheets. A potential measurement of surface roughness and katabatic wind intensity. J. Glaciol., 36, 133-142.

Rémy F., B. Legrésy, S. Bleuzen, P. Vincent, and J.F. Minster, 1996, Dual-frequency Topex altimeter observation above Greenland. J. of Electromagnetic waves and application. vol. 10, 1505-1523.

Ridley, J.K. and K.C. Partington, 1988. A model of satellite radar altimeter return from ice sheets. Int. J. Remote Sensing, 9, 601-624.

Rott, H., H. Miller, K. Sturm and W. Rack, 1994. Application of ERS1 SAR and Scatterometer data for studies of the Antarctic Ice Sheet. Proceedings Second ERS1 Symposium. Space and Environment, Hamburg, ESA SP-361.

Surdyk, S. and M. Fily, 1993. Comparison of the passive microwave spectral signature of the antarctic ice sheet with ground traverse data. Annals of Glaciol. 17, 161-166.

Ulaby, F.T., Moore R.K., and A.K. Fung, 1982. Microwave Remote Sensing, Vol.3, Addison-Wesley Publishing Company, 1100 pp.