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Penetration depth of interferometric synthetic-aperture radar signals in snow and ice

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Abstract. Digital elevation models of glaciated terrain produced by the NASA/Jet Propulsion Laboratory (JPL) airborne interferometric synthetic-aperture radar (InSAR) instrument in Greenland and Alaska at the C- (5.6 cm wavelength) and L-band (24-cm) frequencies were compared with surface elevation measured from airborne laser altimetry to estimate the phase center of the interferometric depth, or penetration depth, δ_p. On cold polar firn at Greenland summit, δ_p = 9±2m at C- and 14±4m at L-band. On the exposed ice surface of Jakobshavn Isbrae, west Greenland, δ_p = 1±2 m at C- and 3±3 m at L-band except on smooth, marginal ice where δ_p = 15±5 m. On colder marginal ice of northeast Greenland, δ_p reaches 60 to 120 m at L-band. On the temperate ice of Brady Glacier, Alaska, δ_p is 4±2 m at C- and 12±6 m at L-band, with little dependence on snow/ice conditions. The implications of the results on the scientific use of InSAR data over snow/ice terrain is discussed.

1. Introduction

Topographic information is essential to study mountain glaciers and ice sheets. Interferometric synthetic-aperture radar (InSAR) is a powerful technique for mapping snow/ice topography, at a high spatial resolution, over large areas, independent of solar illumination, cloud cover, and the presence of recognizable surface features. Several airborne systems were developed and flown in the 1990's to produce digital topographic maps from radar interferometry. These efforts culminated in the Shuttle Radar Mapping (SRTM) mission in early 2000 which mapped the topography of the Earth between ±56° latitude at C-band frequency (5.6 cm wavelength), 30-m posting and 10-m vertical precision (http://www.jpl.nasa.gov/srtm/).

Despite numerous technological and scientific advances, we do not know how deep into snow/firn/ice InSAR signals interact, and how penetration depth (or phase center of the scattering volume), δ_p, may vary with radar frequency, radar imaging conditions, and the physical and electrical characteristics of snow/firn and ice. The results have in turn important implications on the scientific use of the radar data, and the design and performance estimation of future radar systems.

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To evaluate the performance of InSAR topographic mapping over snow/firn and ice, the NASA/JPL Topsar instrument (http://airsar.jpl.nasa.gov/) was deployed in Greenland in May 1995 and Alaska in June 1996 to survey outlet glaciers and inland ice areas of varying snow/firn and ice conditions. Data collection was coordinated with overflights by the NASA/Wallops airborne laser altimeter in Greenland [*Krabill* et al., 1999], and the University of Alaska airborne laser altimeter system in Alaska [*Echelmeyer* et al., 1996]. Laser altimetry provides an accurate surface reference to evaluate the performance of InSAR topographic mapping.

2. Observations

The NASA/JPL Topsar instrument operates on-board the NASA DC-8 aircraft from a flying altitude of 10 km, simultaneously at the C- (5.6 cm wavelength), and L- band (24 cm) frequencies. The C- and L-band topographic maps are referenced to the WGS84 ellipsoid with a 10-m posting using kinematic Global Positioning System (GPS), inertial navigation data, and ground control points (GCP). GCPs were selected in areas of known elevation with no radar penetration such as ocean water, rock outcrop or building structures. The root-mean-square elevation accuracy is 2-3 m at C- and 4-6 m at L-band.

The NASA/Wallops ATM (Airborne Topographic Mapper) is an airborne laser altimetry system developed to measure elevation changes of the Greenland ice sheet. It uses a conical-scanning device on-board a P-3 aircraft, whose location is determined by kinematic GPS techniques. The flying altitude is 400 m, the swath width is 140 m, each laser foot print is 1 m, and the root-mean-square elevation accuracy is 10 cm [*Krabill* et al., 1999].

A similar airborne laser system was developed by the University of Alaska [*Echelmeyer* et al., 1996] to measure volume changes of mountain glaciers in the north-western coastal regions of North America. The system flew in a single engine aircraft aircraft along 30 mountain glaciers of Alaska, Washington and British Columbia in 1995 and 30 glaciers in 1996. Its vertical accuracy is 20 cm [*Adalgeirsdottir* et al., 1998].

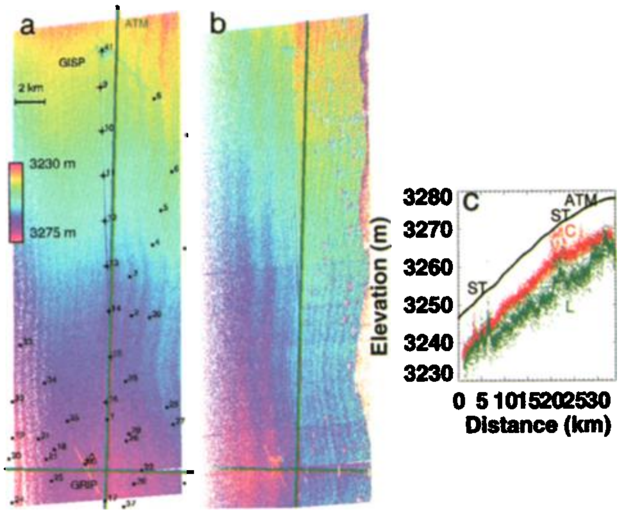

Figure 1. Greenland Summit, 05-1995. (a) C- and (b) L-band topography overlay on radar brightness. Ripples in (b) (±2-3 m) are radar-processing artefacts. White spotches in far range of (b) are spurious echoes from a coastal radar. (c) elevation from laser (black line), C- (red) and L-band (green) data. ST = snow tracks.

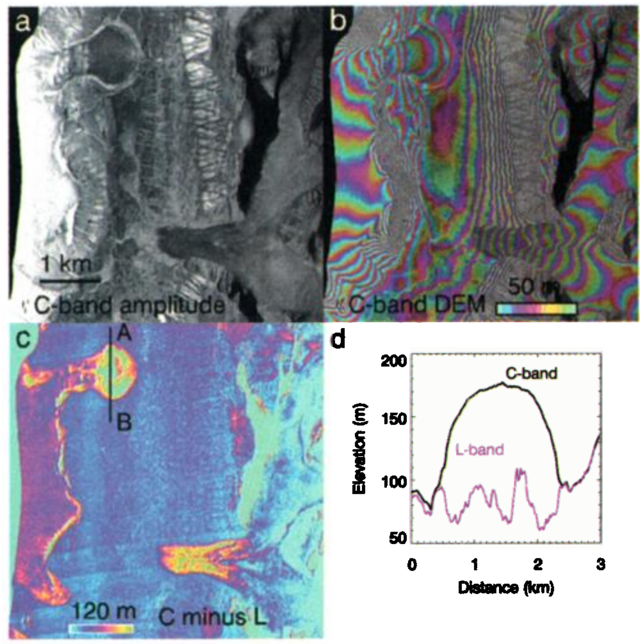

Figure 3. Northeast Greenland piemond glaciers, 05-1995. (a) C-band radar brightness; (b) C-band topography; (c) C- minus L-band elevation. (d) C- (black) and L-band (red) elevation along A-B in (c)

3. Results

Greenland summit exhibits some of the driest snow conditions. In the Topsar imagery acquired on May 18, 1995, the GISP (72.579N, 38.458W, 3255 m) and GRIP (72.576N, 37.627W, 3279 m) drilling facilities and snow tracks in between the two sites appear radar bright (Fig.1a-b), standing several metres above the surrounding surface. ATM data from 1995 and GPS points from 1991, 1992 and 1993 (K. Keller, pers. comm., 1997) were used as ground control points along the snow tracks and drilling sites where radar penetration was presumed to be negligible. The GPS survey points were corrected for a 1.2 m systematic bias with ATM attributed to differences in year of acquisition and reference base station. Penetration depth is 9±2 m at C- (Fig. 1c) and 14±4m at L-band. No significant variation in δ_p (±1-2m) is detected across the swath.

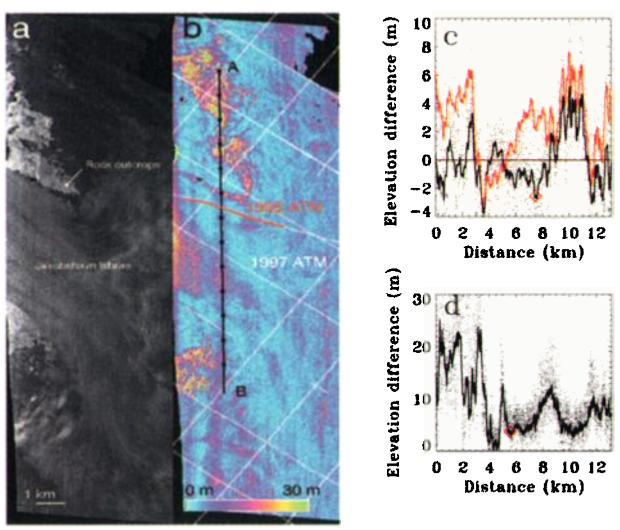

Figure 2. Jakobshavn Isbrae, 05-1995. Glacier flow is from right to left. (a) C-band radar brightness. (b) C- minus L-band elevation, with 1995 ATM flight lines in red, and 1997 ATM in white. (c) Laser minus C- (black) and L-band (red) elevation along 1995 ATM from left to right in (b). (d) C- minus L-band elevation along A-B in (b). Red diamonds in (c) and (d) show intercept between A-B and ATM 1995.

Jakobshavn Isbrae (69.15N, 48.31W) is a major outlet glacier of the west coast of Greenland surveyed by ATM on May 18, 1995 (red line in Fig. 2a) and May 15, 1997 (white line in Fig. 2a). The ATM 1997 data collected on rock outcrops were used as GCPs. Up the center of the glacier (Fig. 2c), δ_p is 1±2 m at C- and 3±3 m at L-band. The penetration differential between C- and L-band, however, reaches 20 m near the ice margin (profile A-B in Fig. 2d) and 10 m in the more crevassed central part of glacier. Comparison of the 1995 and 1997 ATM reveals a mean glacier thinning of 3.4 m in two years. Applying this correction to the 1997 data and comparing the results with InSAR suggests no penetration ±1-2 m of C-band signals over the marginal areas of exposed, slow-moving, marginal ice.

The third site is located at 79.9N, 19.3W, in the northeast corner of Greenland, near Nioghalvfjerdbrae glacier [*Rignot* et al., 2000], in a region of dry and cold climate. Glacier piedmont lobes of circular shape develop along the ice sheet margin, with flow speed typically less than 100 m/yr (Fig. 3). The glacier surface was dry, smooth and snow free in its lower reaches at the time of the overflight (May 1995). No laser altimetry data was collected in that sector. The comparison of C- and L-band reveals a large difference in δ_p, up to 120 m at the lowest ice elevation (Fig. 3d). The pronounced topographic variations in L-band topography along A-B, which do not appear in the C-band data, probably reveal interactions with the bedrock underneath the glacier tongue. Hence, L-band radar signals may penetrate the entire ice substrate in this area.

Brady Glacier, Alaska (58.50N, 136.75W) and Reid Glac -ier (58.75N, 136.75W) to the north are two small coastal, temperate glaciers that drain ice into the Gulf of Alaska. This region was surveyed by laser altimetry on June 5, 1995 (Fig. 3a) and June 12, 1996 by Topsar. The ocean surface was used as a control for the Topsar topography. At this time of the year, ice was exposed from sea level to about 800 m elevation. At 1500 m elevation, snow was probably damped. Penetration depth is 5±3 m at C- and 10±4 m at L-band. No particular correlation was found between δ_p and snow/ice conditions although spatial variations in δ_p are larger at low elevation, over exposed ice.

Comparison of the 1995 laser data with 1950's topography, and laser profiles from late May 2000 indicate a mean thinning rate in the mid 1990's of 0.97 m/yr for these glaciers (Echelmeyer and Harrison, unpublished data). Thinning varies with elevation from 3 m/yr at sea level down to 0.3 m/yr at 1000 m elevation. If we use these thinning rates to convert the 1995 laser data into an 1996 equivalent, δ_p becomes 0±3 m at low elevations at C-band and 4±3 m at high elevation. At L-band, δ_p is 7±4 m.

The last test site is Columbia Glacier, Alaska (61N, 147W), a major coastal glacier of Alaska which has retreated rapidly in recent years [*Meier* et al., 1994]. The glacier was surveyed by Topsar on June 12, 1996 and on June 1, 1994 by laser altimetry along the line shown in Fig. 5a-b. Topographic control was provided by the ocean surface. No L-band topography was available at that site. The results show δ_p varying 40 to 10 m from the glacier head to the glacier upper reaches. Based on the results described earlier indicating little penetration of C-band signals over exposed ice, we attribute much of the difference in elevation to glacier thinning between 1994 and 1996.

The red curve in Fig. 5c shows the glacier thinning measured using laser data from May 1994 to May 1999 (Echelmeyer and Harrison, unpublished data): 28 m/yr near the terminus (sea level to 100 m elevation), 10 m/yr at 1000 m elevation, and 2 m/yr at 2000 m elevation. An independent estimate of glacier thinning obtained from aerial photography (B. Krimmel, pers. comm. 1999) indicates a thinning rate of 22 m/yr at sea level in 1994-1996 decreasing to 0 m/yr at 1400 m where the photographic method becomes less reliable. These results support the hypothesis that penetration depth is less than 1-2 m at C-band. They also illustrate the usefulness of C-band topography (e.g. SRTM) for measuring changes in glacier volume.

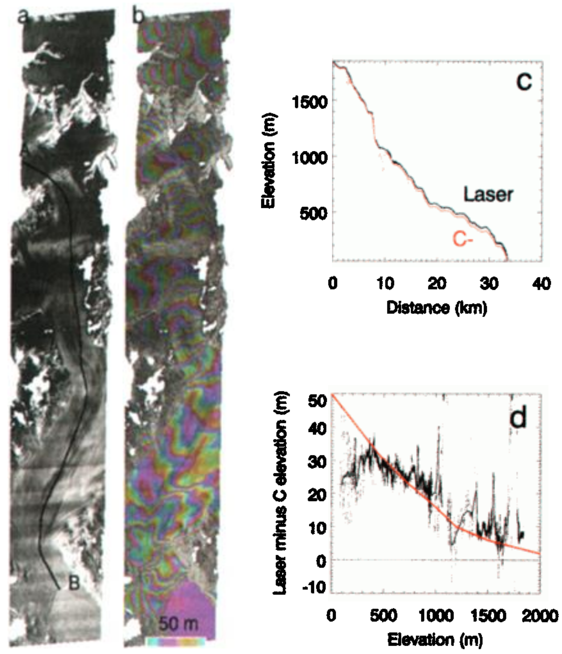

Figure 5. BColumbia Glacier, Alaska, 06-1996. (a) C-band radar brightness; (b) C-band topography; (c) elevation along A-B in (a) from laser (black) and C-band (red). (d) Laser minus C-band elevation along A-B. The red curve shows the amount of glacier thinning inferred from an independent study.

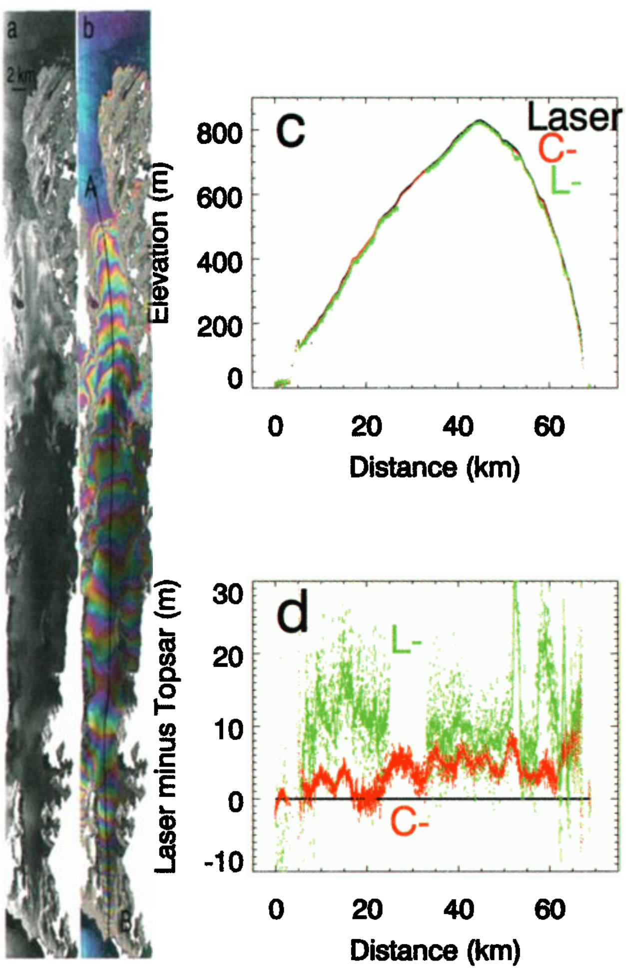

Figure 4. Brady Glacier, Alaska, 06-1996. (a) C-band radar brightness; (b) C-band topography (50-m contour); (c) elevation along A-B shown in (a) from laser (black), C- (red) and L-band (green); (d) laser minus C- and L-band elevation along A-B.

Conclusions

The study illustrates that InSAR topography generally closely follows the surface profile of ice sheets and glaciers (Bindschadler et al., 1999), yet the phase center of the radar data is typically located several meters below the surface. C-band penetration is small (1-2 m) on exposed ice, but up to 10 m on dry, cold firn. This depth of penetration is reasonable enough for most glaciological applications, including the measurement of volume changes. Penetration depth is 5-10 m greater at L-band, and up to 60-120 m on smooth, cold, exposed ice. This extreme level of penetration is only expected over a small fraction of the world's glaciated terrain, yet it demonstrates that L-band penetration can significantly bias the interpretability of the radar data.

Long-wavelength radars, however, present important advantages. Deep penetration is preferable for velocity mapping in order to maintain temporal stability of the radar signals over a long time period, as demonstrated by the SIR-C results over Patagonia ice [*Rignot*, et al., 1996; *Michel and Rignot*, 1999], and also to measure strain rates deeper into snow and ice. Monitoring changes in glacier volume is probably best addressed by short-wavelength radars, but long-wavelength systems are probably required for repeat-pass interferometry over warm ice.

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