Strong spatial variability of snow accumulation observed with helicopter-borne GPR on two adjacent Alpine glaciers

H. Machguth,[1] O. Eisen,[2,3] F. Paul,[1] and M. Hoelzle[1]

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[1] This study compares high-resolution helicopter-borne radar measurements to extensive ground-based profiling of the snow cover on Findel- and Adler Glacier, Switzerland. The results demonstrate that derived accumulation values of either method are well in accordance. The spatial distribution of radar based snow depth allows a clear distinction of three zones of different accumulation characteristics: (1) The lower part of Findel Glacier shows a clear altitudinal trend while (2) the upper part has no trend in altitude but high spatial fluctuations in snow depth. (3) Adler Glacier's accumulation characteristics are similar to zone (2). However, despite their close vicinity, accumulation on (3) is reduced by 40% compared to (2). The observed strong spatial variability emphasises the need for spatially continuous measurements for studies involving accumulation on glaciers. Finally, reasons for observed variations (e.g., preferential snow deposition and snow redistribution) are discussed.
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1. Introduction

[2] Utilisation of radio-echo sounding techniques to determine past and present accumulation rates has become a standard method, especially in polar glaciology. On polar ice sheets and polythermal glaciers this method is rewarding because of the low absorption of electromagnetic energy in cold firn. On temperated glaciers, application is usually restricted to sounding of ice thicknesses with low frequencies (<100 MHz). High-resolution measurements are mostly performed with so-called ground-penetrating radar systems (GPR) [e.g., *Richardson et al.*, 1997; *Kohler et al.*, 1997], operating directly from the surface. They are thus relatively time consuming and the spatial coverage is limited to accessible areas. Application of high-resolution airborne radar, capable of mapping annual accumulation, is still rare and mostly limited to fixed-wing aircraft [e.g., *Kanagaratnam et al.*, 2004]. To cope with the difficulties imposed by measurements in mountainous terrain and valley glaciers, airborne radar is most suitable. Apart from quasi-airborne measurements from an aerial tramway [*Yankielun et al.*, 2004], helicopter-borne radar provides the most reasonable platform and has been applied for studies of sea and river ice [*Wadhams et al.*, 1987; *Arcone and Delaney*, 1987; *Melcher et al.*, 2002], snow on ground [*Marchand et al.*, 2003] or glacier thickness measurements [*Thorning and Hansen*, 1987; *Damm*, 2004]. However, on mountain glaciers only few studies use helicopter-borne radar to investigate the properties of the snow cover and none was so far dedicated to the spatial accumulation distribution. For instance, *Arcone and Yankielun* [2000] focus on intraglacial features in the ablation zone of a glacier, whereas *Arcone* [2002] investigates processing techniques and autonomously derives physical properties of temperate firn.

[3] Today's glacier mass-balance models are based on sophisticated formulations of the energy fluxes while accumulation processes are treated in a very simple way: precipitation varies only with altitude and any processes of snow redistribution are neglected [*Hock*, 2005]. Using a similar model (described by *Machguth et al.* [2006]) we have calculated the mass balance distribution for a glacierized catchment in southwestern Switzerland, including Findel Glacier (length 7.2 km, area 15.3 km^2) and its neighbour Adler Glacier (3.1 km, 2.0 km^2) (Figure 1). (In this paper we refer to "Findel Glacier". According to the maps of the Swiss Federal Office for Topography (swisstopo) this is the official name of the glacier. However, in most glacier inventories the glacier is called Findelen Glacier.) The modelling for the 1971–1990 time period resulted in a very positive mass balance for Adler Glacier of 0.7 m water equivalent (m we) while its larger neighbour's mass balance was negative (−0.25 m we). In fact, the shrinkage of both glaciers indicates that a very positive mass balance is unrealistic. We assume that, in reality, accumulation on Adler- and Findel Glacier differs strongly. The model's failure is most probably caused by treating precipitation as a function of altitude alone and by ignoring snow redistribution.

[4] In this study we combine high-resolution helicopter-borne GPR measurements and extensive ground-based profiling of the snow cover to determine the spatial distribution of accumulation and to validate our assumption of a strong local variability in accumulation. In contrast to methods that require measurements at two points in time (e.g., mapping of elevation changes with laser altimetry, accumulation stakes without snowpits), GPR has the strong advantage that changes over time in surface elevation (e.g., melt, settlement of the underlying snow cover and glacier movement after the first measurement) do not affect the accuracy of the measurements. Furthermore, measuring only once requires less logistical efforts.

2. Methods

2.1. Radar System and Data Acquisition

[5] A commercial Noggin Plus 500 radar (Sensors & Software Inc., Canada,) was operated with a shielded

[1] Geographisches Institut, Universität Zürich, Zurich, Switzerland.
[2] Versuchsanstalt für Wasserbau, Hydrologie und Glaziologie, ETH Zürich, Zurich, Switzerland.
[3] Alfred-Wegener-Institut für Polar- und Meeresforschung, Bremerhaven, Germany.

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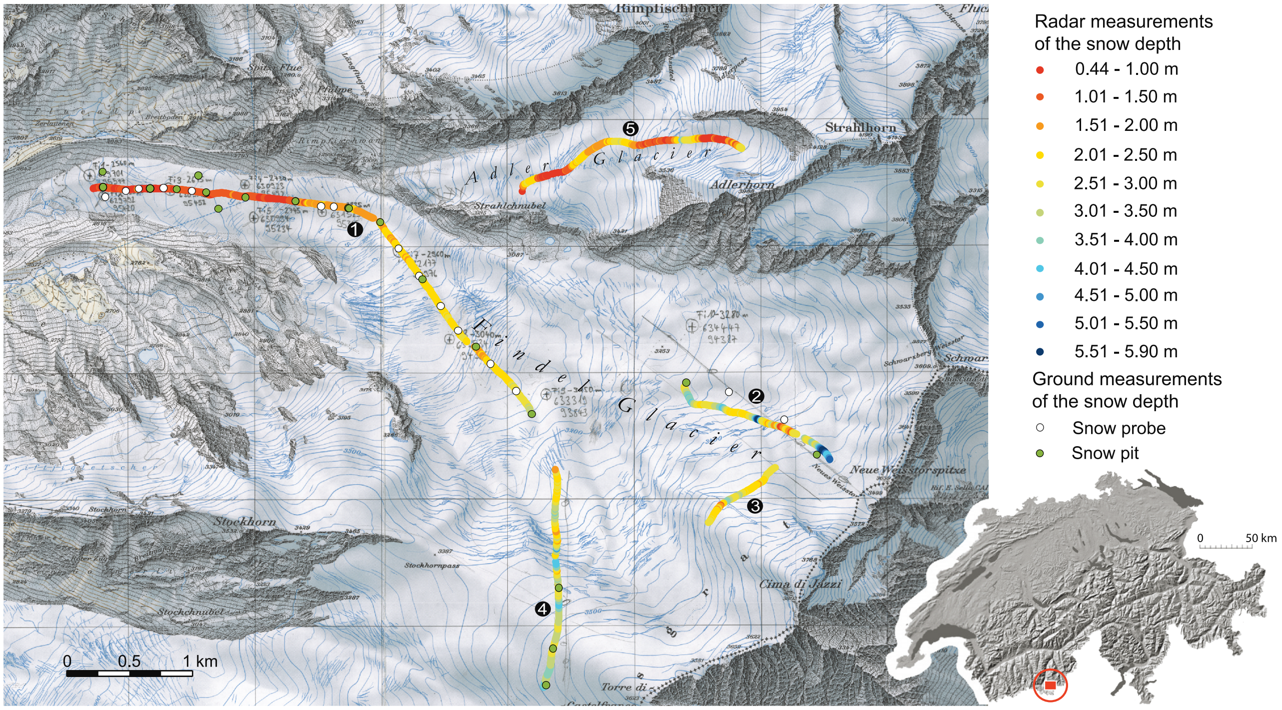

Figure 1. Map of Findel and Adler Glacier (red square in inset on Swiss map) with radar profiles (black numbered circles) and ground measurements. Color indicates radar-based snow depth. Reproduction of the background map with permission from swisstopo (BA067878).

antenna (15 cm transmitter-receiver separation) at a centre frequency of 500 MHz and 400 MHz bandwidth. The helicopter–radar combination was developed by Airborne Scan, Visp, Switzerland. The antenna was mounted to the helicopter runner. Data acquisition was performed in a constant-time triggering mode, with a time increment of 0.1 s between traces. Helicopter speed was about 6 m/s on average, resulting in a mean trace increment of 0.6 m. Helicopter altitude above ground was between 2.5 and 30 m, being 11 m on average, resulting in a footprint size in the order of a few tens of m^2. 1876 samples were recorded per trace with a 0.2 ns sampling interval, resulting in a 375 ns time window. Four-fold pre-storage stacking of traces at a pulse repetition frequency of 25 kHz was applied. A GPS antenna was mounted at the nose of the helicopter. For each recorded trace the GPS-position was simultaneously stored. Real-time GPS results in a position accuracy of <10 m.

[6] The radar flight was accomplished on 9 May 2005, ground measurements for validation were conducted on 6 and 7 May 2005. During all four days air temperature at the glacier terminus was slightly below the freezing point. Below 2700 m a.s.l. the snow was wet and some water drained off at the snow-ice interface. Above 3200 m the entire snow pack was dry. In approximately 25 minutes of flight 10.0 km of radar profiles were collected, thereof 1.9 km on Adler Glacier (Figure 1). On Findel Glacier radar profiles reach from 2570 to 3560 m a.s.l., on Adler Glacier they reach from 3240 to 3690 m a.s.l. Under the favourable weather conditions the sites of snow pits and -probes were visible from the helicopter allowing a consecutive visual determination of the flight direction. According to GPS data, the helicopter passed 26 snow pits and probes at a distance of <5 m, four at 5–15 m and two at 15–30 m. Two snow probes and one snow pit were missed by 60–130 m because they could not be found again.

2.2. Radar Data Processing

[7] Post-recording processing and radar-data analyses were carried out with ReflexWin (Sandmeier Scientific Software, Germany). Processing steps include dewowing (high-pass filter), background removal, application of a gain function to mainly compensate for spherical spreading, and additional filtering. The varying helicopter altitude above ground required a static correction of each trace to the first break of the surface reflection (time-zero correction). Due to the relatively smooth surface topography and small layer thickness compared to the helicopter altitude no migration was necessary. Along most profiles one or more distinct reflections of different magnitude are visible (see sample radar profile (Figure 2 or Figure S1 in auxiliary material[1]). Tracking of the uppermost strong continuous reflection horizon resulted in continuous profiles of last winters snow layer thickness (Figure 1). No interpretation was performed where no distinct reflection horizon was visible. Density measurements in the snow pits (see below) yielded a mean density of 400 kg m^{-3}. Based on the linear and quadratic conversion formulas of *Tiuri et al.* [1984] and *Kovacs et al.* [1995], respectively, conversion of the radar data from time to depth domain is carried out with a mean wave speed in snow of 2.2 10^8 ms^{-1}. Using the same mean density of 400 kg m^{-3} the layer thickness is converted to water equivalent.

[1]Auxiliary material is available in the HTML.

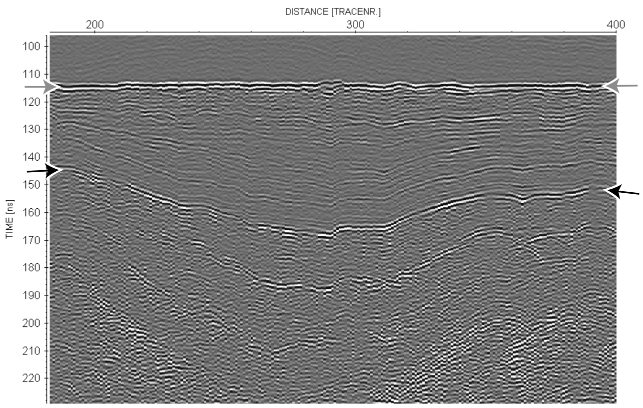

Figure 2. A section of the radar profile 2 at 3460 m a.s.l. The horizontal axis corresponds to approximately 130 m. The varying snow depth, the internal layering of the snow pack as well as previous years' firn layers with some internal structures are clearly recognizable. Grey arrows indicate the air-snow interface and black arrows indicate the snow-firn interface.

2.3. Snow Pits and Snow Probes

[8] Snow pits and snow probes were used to measure snow depth and density from the actual snow surface down to the ice surface. Where the snow cover outlasted the previous melting season (the remaining accumulation zone of 2004 is located above 3300 m a.s.l. on Findel Glacier, for Adler Glacier no data is available), snow pits were dug down to the horizon of the previous autumn's snow surface. According to data from the meteo station Gornergrat (located 5 km west of Findel Glacier at 3100 m a.s.l.), the long melt season of 2004 ended with heavy snow falls on 10 October. Consequently, the snow depth measured with radar, snow pits and -probes was accumulated within the time span of about 10 October 2004 to 7 May 2005. Coordinates of snow pits and -probes were captured with a hand-held GPS (position accuracy of 5–15 m). Within an altitudinal range from 2590 to 3510 m a.s.l. 20 snow pits have been sampled and snow depth and density as well as the internal layering of the snow pack was determined in all of them. This sample size was enhanced by 19 measurements of snow depth with snow probes. However, defining previous autumn's snow surface using snow probes is sometimes difficult (e.g., misinterpretations due to ice layers within the winter snow pack). Consequently, snow probes were restricted to the ablation area and for every test site the mean value of nine snow probes, sampled within a radius of 7 m, was calculated.

2.4. Data Merging

[9] Field measurements were only used to validate the interpretation of the radar profiles, analyses were conducted separately. Neither GPS data nor a map were used for the interpretation of the radar profiles, thus the interpreter's knowledge about the field measurements could not influence his interpretation of the profiles. The data sets were joined within a GIS software (ArcGIS 9.1). The twenty closest radar traces to every ground measurement were selected and their mean value is used below for comparison with the corresponding ground measurement.

3. Results

[10] The transition between winter snow and ice or winter snow and snow having outlasted the previous summer is in general clearly recognizable in the radar profiles. A total of 0.6 km radar profiles (6% of the total length) did not allow any tracking due to lacking or disturbed layering. Most of these zones are located within crevassed areas. Further analysis of the data is based on almost 15000 radar traces, representing the remaining 9.4 km of radar profiles. The thickness of the winter's snow layer varied in the ground measurements (radar measurements) from 0.32 (0.44) to 4.4 (5.9) m. The specific density measured in the snow pits is not correlated with altitude and varies from 360 to 470 kg m^{-3}. The mean density is 400 kg m^{-3} with a standard deviation of 30 kg m^{-3}. Figure 3 shows the agreement between radar profiles and all snow pits and -probes where the horizontal distance to the radar track is less than 30 m. The linear regression yields a correlation coefficient of R^2 = 0.84. Three data points must be considered outliers, as discussed below. Excluding these data points yields R^2 = 0.96.

[11] According to the radar profiles three zones of different accumulation characteristics can be distinguished: the lower part (profile 1) and the upper part of Findel glacier (profiles 2, 3 and 4), as well as Adler Glacier (profile 5). The lower part of Findel Glacier shows a clear correlation between altitude and snow cover thickness (R^2 = 0.81) and the fluctuations in snow depth are small (Figure 4). On Adler Glacier and the upper part of Findel Glacier accumulation has no altitudinal trend ($R^2 \leq 0.01$). Fluctuations in snow depth are very large. The correlation coefficient

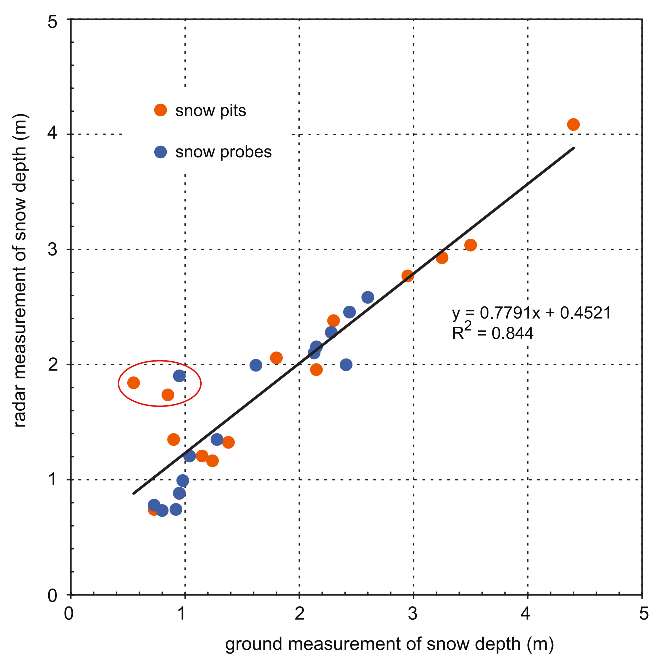

Figure 3. Comparison of radar- and ground-based measurements of snow depth. The three outliers are marked with a red circle.

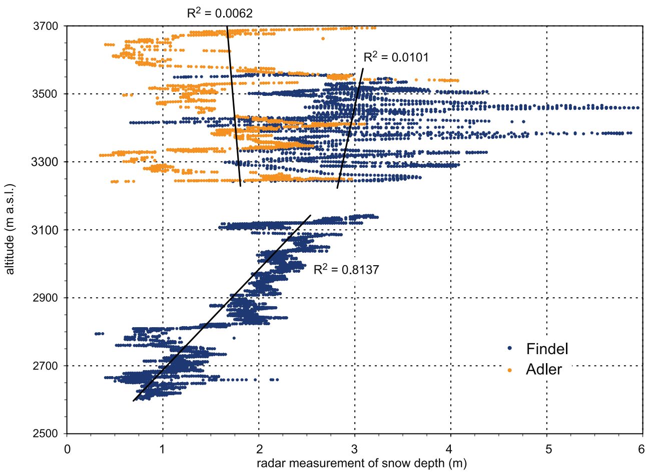

Figure 4. Snow depth versus altitude for all profiles on Findel and Adler Glacier.

calculated for the upper part of Findel Glacier is based on all traces of the profiles 2, 3 and 4 but also represents well the characteristics of every individual profile: none shows any significant trend with altitude; and all show large fluctuations of snow depth. At altitudes where radar profiles exist on both glaciers (3240 to 3560 m a.s.l.) the average accumulation is 2.98 m snow (5704 traces) on Findel and 1.80 m snow (2164 traces) on Adler Glacier. The accumulation on Adler Glacier is 0.5 m we or 40% lower than on Findel Glacier.

4. Discussion

[12] Our correlation between ground and radar measurements confirms the results of earlier studies with helicopter-borne radar in non glacierized landscapes, having achieved correlation coefficients of 0.82–0.97 [e.g., *Marchand et al.*, 2003]. The three outliers mentioned above are presumably caused by misinterpretations in the field. Snow-probe measurements at those three sites (see Figure S2 in auxiliary material), located within a diameter of 100 m at 2850 m a.s.l. are contradictory (0.5–1.7 m snow). A massive ice layer at 0.5 m depth was recorded in the snow profile closest to them. We assume that this layer was interpreted as the ice surface. The layer is visible in the radar profile as a long, massive and smooth reflection horizon at a depth of 0.5 to 1 m. Normally, the snow-ice interface appears more uneven because of previous surface melt and is similar to the second, less distinct reflection horizon at about 1.8 m in that profile. The example shows that in situ measurements might present only partial truth. A correct interpretation of the internal layering of the snow pack is essential for both methods. Consequently, mutual comparison is important to combine their strengths. In the ablation area snow pits allow precise point measurements while the large footprint of helicopter-borne GPR has the advantage of averaging out the bumpiness of the ice surface. In the accumulation area, both methods have to deal with the difficulty to correctly identify last autumn's snow horizon.

[13] The overall pattern of accumulation distribution on Findel Glacier corresponds well to the observations based on snow probes on other alpine glaciers (e.g., Vernagtferner in Austria, *Plattner et al.* [2006]). On the lower part of Findel Glacier the correlation between altitude and snow depth can partly be explained by melt during long warm weather periods in March and April 2005. The large deviations in accumulation between Findel and Adler Glacier are very unlikely to be caused by enhanced melt in March and April due to differing surface exposition: assuming that the entire difference in global radiation due to exposition (20 W m^{-2}) is available for melt and considering the high snow albedo of approximately 0.8 for this period, melt on Adler Glacier is estimated to be 0.05 m we higher than on Findel. We therefore assume that the large deviations in accumulation are primarily due to spatial variability of precipitation and redistribution of snow by wind. The area of investigation receives high amounts of precipitation under southerly wind directions. Reduced wind speed leeward from ridges results in enhanced precipitation and additional snow deposition from wind transport [e.g., *Föhn and Meister*, 1983; *Gauer*, 2001]. Findel glacier is located directly leeward of the main ridge and profits from this effect, whereas Adler Glacier is farther away. In addition, strongly reduced accumulation is also observed in crevassed zones, probably caused by topography and microscale turbulences. Within this study we have only assessed the spatial and not the temporal variability of accumulation. However, taking into account the concurrent shrinkage of both glaciers, we assume that the observed deviations in accumulation of both glaciers for the winter 2004/05 are not exceptional.

5. Conclusions and Outlook

[14] The data of both methods used in this study are in very good agreement. The observed distribution of the snow cover confirms our assumption of strongly reduced accumulation on Adler Glacier. Our results emphasize that the distribution of accumulation is not simply a function of altitude, confirming other studies [e.g., *Winstral et al.*, 2002]. We suppose that the spatial variability of precipitation as well as the redistribution of snow are mainly governing the accumulation distribution. The observed variability of the mass balance for the accumulation period (0.5 m we) is one order of magnitude higher than the error of melt calculations for the entire ablation period with energy balance models [compare *Arnold et al.*, 1996; *Obleitner and Lehning*, 2004]. The results underline that major improvements in glacier mass balance modelling can be achieved by focusing on the accumulation processes. Helicopter-borne GPR is recommended as a reliable tool for time-saving and accurate mapping of the snow cover. The method allows to enhance the sparse database on accumulation distribution toward both spatial and temporal variability, providing a sound data basis for glacier monitoring (e.g., yearly repeated measurement of winter balance), for statistical analyses (e.g., digital elevation model attributes) or for validation and calibration of physical modelling. Field measurements remain essential for mutual validation and to determine snow density.

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O. Eisen, Laboratory of Hydraulics, Hydrology, and Glaciology, ETH Zürich, CH-8092 Zurich, Switzerland.

M. Hoelzle, H. Machguth, and F. Paul, Physical Geography Division, Department of Geography, University of Zurich, Winterthurerstr. 190, CH-8057 Zurich, Switzerland. (hoelzle@geo.unizh.ch; machguth@geo.unizh.ch)