The Cryosphere Discussions

Spatially extensive estimates in annual accumulation in the dry zone of the Greenland Ice Sheet inferred from radar altimetry

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Abstract

We present estimates of accumulation rate along a 200 km transect ranging in elevation from 2750 to 3165 m in the dry snow zone on the western slope of the Greenland Ice Sheet. An airborne radar altimeter is used to estimate the thickness of annual internal layers and, in conjunction with ground based snow/firn density profiles, annual accumulation rates between 1998 and 2003 are derived. The observed inter-annual variability is high, with the annual mean accumulation rate estimated at 0.372 m.w.e. yr^{-1} (s.d. ± 0.063 m.w.e. yr^{-1}). Mean accumulation rates modelled using meteorological data overestimate our results by 13% on average, but by 37% and 32% in the years 2001 and 2002. The methodology presented here demonstrates the potential to obtain accurate and spatially extensive accumulation rates from radar altimeters in regions of ice sheets where field observations are sparse.

1 Introduction

It is known that the arctic region is warming and that the margins of the Greenland Ice Sheet are experiencing substantial thinning (Hassol, 2004; Steffen et al., 2004; Zwally et al., 2005; Pritchard et al., 2009). Increasing fresh water input from the Greenland Ice Sheet (GrIS) to the oceans has created a growing concern about the effects it could have on climate and sea level rise. Although changes in the volume of an ice sheet may be estimated by observing variations in its surface elevation (e.g. Wingham et al., 1998; Shepherd et al., 2001 and 2002; Zwally et al., 2005) and from gravity measurements (e.g. Velicogna, 2009), these methods do not distinguish ice dynamics and accumulation fluctuations which may introduce errors in mass balance estimates.

Observations show that surface melt extent and duration have increased dramatically since the late 1970s (Abdalati and Steffen, 1997; Steffen et al., 2004; Mote, 2007; Hanna et al., 2008). Meanwhile, there are reports of an increase in the surface elevation at higher altitudes since 1990 (Krabill et al., 2000; Nghiem et al., 2005; Hanna et

al., 2006), supporting calibrated climate models which show that some regions of the ice sheet have experienced higher than average accumulation rates, especially in the south (Burgess et al., 2010). A recent study reveals that the total accumulation in the GrIS is up by 24%, and that the surface mass balance estimates to be 63% higher than

5 previous calculations (Ettema et al, 2009). However, Greenland's weather and accumulation patterns have high spatial and temporal variability (Mosley-Thompson et al., 2001; Steffen et al., 2004; Wang et al., 2007), and elevation changes may be caused by factors other than accumulation and melt such as snow compaction and densification processes (Parry et al., 2007). For this reason, accurate estimates of mass

10 balance made from elevation change observations require continuous measurements of accumulation and compaction rates across different regions of an ice sheet. Direct observations of snow accumulation in Greenland are limited, and measurements are affected by wind patterns and for the most part made at stations located in coastal regions (Dethloff et al., 2002), while accumulation estimates derived from ice cores and

15 snow density measurements are also limited spatially.

Here, we use airborne radar altimetry data and snow density measurements derived from snow/firn cores and neutron-probe observations obtained over the dry snow zone of the GrIS to estimate annual accumulation rates for the years 1998-2003 along a 200 km transect ranging in elevation from 2750 m to 3165 m on the western slope

20 of the GrIS (Fig. 1). The estimates obtained are compared with snow accumulation rates derived from a weather prediction numerical model run retrospectively using data from the European Centre for Medium-Range Weather Forecast (ECMWF) 45-year (1957–2002) reanalysis ERA-40 and ECMWF operational analysis for 2002-onwards data (Hanna et al., 2005, 2006 and 2008). This model has been recently corrected and

25 calibrated for Greenland with recent field measurements (Bales et al., 2009). Technical advances in radar altimetry have made the European Space Agency's Airborne SAR/Interferometric Radar Altimeter System (ASIRAS) capable of resolving internal layers in the snowpack and firn (Hawley et al., 2006; Helm et al., 2007). We extend the previous ASIRAS analyses in order to resolve spatial and temporal trends in the

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estimated accumulation rates presented. ASIRAS is designed to emulate the operation of the SAR/Interferometric Radar Altimeter (SIRAL) onboard the recently launched CryoSat-2 (Wingham et al., 2006), thus one of the objectives of this paper is to demonstrate the viability of estimating accumulation rates from CryoSat-2 or other space

5 borne altimeters with similar operating characteristics. CryoSat-2 has a poorer vertical resolution, and due to its larger footprint, it has higher speckle noise, so further data processing would be required for an analysis of accumulation layers.

2 Data and methodology

The data reported here were derived from observations made by ASIRAS during the
10 2004 CryoSat Validation and Calibration field experiment. On September 17th 2004, the ASIRAS radar altimeter was flown along the Expedition Glaciologique au Groenland (EGIG) line (Fischer et al., 1995) over the GrIS operating in High Altitude Mode (HAM). We present a 200 km transect, along the EGIG line, on the western slope of the dry snow region of the ice sheet ranging in elevation from 2750 m to 3165 m. The
15 system is a pulse-width limited, coherent, and phase sensitive radar altimeter that operates in the Ku band, at a central frequency of 13.5 GHz and with a bandwidth of 1 GHz. The system has a footprint of 4.5 m along track (after SAR processing) and the large bandwidth provides high vertical resolution, so it is possible to analyze in detail the structure of the observed snowpack.
20   Radar signals can penetrate snow, and are partially backscattered by changes encountered in both snow density and crystal structure. Thus stratigraphically distinct layers in the snowpack as are commonly formed early in the fall in the dry snow region of the GrIS (Autumn hoar) make it possible to identify widespread internal layers reflecting annual accumulation (Hawley et al., 2006). An advantage in using the Autumn
25 hoar to identify annual layers is that it coincides with the commonly used concept of a mass balance year for an ice sheet, lasting typically from the onset of accumulation (in Autumn) in one year to the cessation of ablation (end-of-summer) the next year (Anklin

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and Stauffer, 1994). Thus, when referring to the observed accumulation rate in 1999, we are referring to the period covering ∼October 1998 to ∼September 1999.

To estimate the distance travelled, we calculate the speed of light in snow based on the dielectric permittivity of snow. By neglecting any water content (dry snow), we assume we are working on a lossless media so we can obtain the permittivity of snow n_e directly from the snow density, for which we use mean values for each of the layers detected based on snow density profiles obtained from snow/firn cores and neutron-probe observations. The permittivity for dry snow with a density under $0.5\,\mathrm{kg\,m^{-3}}$ is given by (Yankielun et al., 2004):

$$n_e = (1 + 1.9\rho_s)^{1/2} \tag{1}$$

where ρ_s is the density of snow. Using Eq. 1, we can estimate the speed of light v_c through the snowpack from its linear relationship with the electric permittivity of snow, $v_c = c/n_e$, where c is the speed of light in vacuum. The snow density measurements used to estimate dielectric permittivity and accumulation were obtained from snow/firn cores and neutron-probe observations acquired at Summit, Greenland at 3200 m (Hawley et al. 2008) and at a site located 25 km west of the start of the transect at an altitude of 2650 m (Hawley et al., 2006). Summit is located 150 km north from the surveyed path, but along the ice divide and at only 35 m above the elevation of our easternmost observation. We assume a linear decrease in density with elevation from the lower site to Summit and use the estimated densities at each point along the transect to derive dielectric permittivity and accumulation. Although this is a simplistic approximation, conditions at regions of the GrIS above 2000m remain very similar and there is no reason to suggest that changes in density in this area are not directly dependant on elevation.

We estimated the depth of each observed layer at 25 km intervals along the 200 km transect. At each point, 100 individual waveforms are averaged to minimize small scale variations in layer thickness. There are clear differences in the degree of variability in waveform structure with elevation. Between 2800 m and 2950 m elevation (50 to

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110 km along the transect), the depth of the observed layers vary as much as ± 9 cm between the different waveforms, while at higher elevations no differences larger than the system range accuracy, which is estimated to fall within 3 cm (Hawley et al., 2006), are observed. In a previous study, winter accumulation rate estimated from ASIRAS at one site over the percolation zone was estimated at 0.63 ± 0.06 m.w.e., which compared favourably with the 0.61 m.w.e. measured in the field (Helm et al., 2007). Uncertainty is introduced from the assumption that each layer delimits exactly one annual accumulation period. In reality, the snow density changes that cause radar reflections are not formed at exactly the same time every year. For example, it is known that the 2002 summer melt season was unusually long with areas experiencing 2 weeks more melt than on average (Steffen et al., 2004), so if temperatures remained unusually high across the whole ice sheet, the formation of the autumn hoar may have been delayed, slightly modifying the length of the period for which the accumulation rate estimates are made. However, this is highly unlikely due to the predominant cold temperatures in the dry snow zone of the GrIS.

3 Results and discussion

The radar altimeter observations along the 200 km transect are shown in Fig. 2. Individual waveforms are aligned with respect to the signal in each waveform estimated to be the surface, which reveals horizontal continuity in the firn structure with clear layers observable to a depth of about 10 m (travel time = 80 ns). A strong return from the surface is evident, with several weaker reflections from the volume of the snowpack along the whole transect. These weaker reflections occur at interfaces within the snowpack where sharp changes in the firn density/structure occur, as described earlier, and represent annual accumulation periods. The distance between the annual layers decreases with increasing elevation; this is expected, since the mean snow accumulation rate increases in a westerly direction away from the ice divide along the EGIG line (Thomas, 2004). Seven distinct layers are clearly visible along the whole transect, with several more appearing at depth at higher elevations in the east where snow densities

are lower. It has been observed elsewhere that radar signals penetrate deeper into the snowpack as snow density decreases, thereby revealing additional layers with increasing elevation and decreasing snow density (Scott et al., 2006). The change in thickness of each layer is clearest in 2004 (the top layer), decreasing from 1.57 m at 2750 m (70.45° N, 43.52° W) to 1.16 m at 3165 m (71.17° N, 37.88° W). The layers underneath reveal a smaller change in thickness, likely due to the fact that they have been subject to compaction for longer. For this reason, only accumulation estimates for layers below the 2004 layers are made.

Figure 3a shows the estimated thickness of the annual layers for 1998–2003. The decrease in layer thickness with increasing elevation is evident in all years, and there is substantial inter-annual variability observed in the rate at which thickness decreases with elevation. A decrease in thickness of 14–18% is observed in 2000 and 2003, 23–27% in 1998 and 2002, and 45% and 50% in 1999 and 2001 respectively. Figure 3b combines the estimated annual layer thickness (Fig. 3a) with derived snow density to estimate the total accumulation for each annual layer down to 1998. There are large inter-annual variations in accumulation ranging from 0.271 ± 0.056 m.w.e. yr^{-1} in 1999 to 0.458 ± 0.045 m.w.e. yr^{-1} in 2000 (Table 1). Accumulation rate decreases from west to east along the 200 km transect reducing by between 14–18% for 2003 and 2000; 30–36% for 2002 and 1998; 50–55% for 2001 and 1999 (percentages calculated from the trend shown by a linear regression estimated for each year, with an average correlation coefficient r^2 of 0.826), confirming that the trajectory of moist air masses follows an easterly path along the western slope of the GrIS in the vicinity of the EGIG line. Furthermore, strong katabatic winds may further contribute to the transfer of snow from higher to lower elevations away from the ice divide. It is noteworthy that while the thickness in the annual layers between the lowest and highest elevations along the transect decreased on average by 25%, the net drop in mean accumulation is estimated to be around 33.8%, clearly indicating that any loss (or gain) in the ice sheet mass may not be completely reflected by proportional changes in surface elevation (Parry et al., 2007).

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We calculated the mean accumulation rates at 25 km intervals along the transect (Table 2) and present these in conjunction with both the mean annual estimates derived from the ERA-40 data-based model output and estimates made from two previous field based studies (Fischer et al., 1994; Anklin and Stauffer, 1994). Comparison with field data was only possible at our lowermost point at 2750 m, with an estimated accumulation rate of 0.461 m.w.e. ± 0.099 compared to 0.423 ± 0.012. In practice, the difference is only related to the accuracy of the system to estimate layer thickness, since both in-situ and ASIRAS derived accumulation estimates are made using the same densities. Our estimates are higher from the previously recorded observations, with an increase in accumulation of 19% at sites above 3000 m, but only 6% higher at sites below 3000 m (which is within our error estimate). The ERA-40 model over-predicts the ASIRAS derived accumulation rates by an average of 13.5%. As stated earlier, there is a clear spatial gradient in accumulation rate, with an annual mean of 0.461 m.w.e. ± 0.099 at 2750 m elevation decreasing to 0.296 m.w.e. ± 0.084 at 3165 m. To prove that the assumption of a linear decreasing density from 2650 m to Summit does not introduce large errors, we made a second estimate of the accumulation rates using a constant density for each annual layer. In the highest point (where the snow density is the lowest and thus the error would be the highest) the estimates were 0.0298 m.w.e. ± 0.014 larger, or roughly 10%.

To further illustrate the importance of monitoring short scale spatial and temporal variability, we compare our results with accumulation rates derived from an ERA-40 data-based meteorological model. The model runs with meteorological data thus do not depend on density measurements. The model was downscaled to 5×5 km grid and calbrated against kriged ice-core data mapped to the same grid (Bales et al., 2009). Figure 4 illustrates the spatial and the temporal variability of the estimates derived both from the ASIRAS observations (Fig. 4a), and the model run retrospectively (Fig. 4b). The different approaches reveal very similar patterns with decreasing snow accumulation from west to east and similar inter-annual variability, with the exception of the year 2002, in which the model predictions overestimate our results by 34%. Modelled

accumulation for a given year may be subject to unusual meteorological conditions that affect the model output. It is known that the GrIS experienced unusually high temperatures in 2002 (Steffen et al., 2004; Wang et al., 2007), which increased precipitation in the coastal regions but may not have significantly enhanced accumulation rates in the higher regions of the ice sheet. Coincidentally, the model results show the highest accumulation rates in 50 years for south-eastern Greenland in 2002 (Hanna et al., 2006).

4   Summary and conclusions

We presented estimates of annual snow accumulation rates derived from observations made by the ASIRAS airborne radar altimeter along a 200 km transect of the dry snow region on the western slope of the Greenland Ice Sheet. Different layers are identified within the snowpack, each associated with the annual snow accumulation for the period 1998–2003. The average annual accumulation is estimated at 0.372 ± 0.063 m.w.e., but the rates exhibit both high spatial and temporal variability, with the mean accumulation rate estimated at 0.271 m.w.e. in 1999 and at 0.458 m.w.e. in 2000. A clear gradient in the thickness of each layer and in the associated estimates of annual accumulation is seen along the transect, with a $34.2\% \pm 16\%$ mean decrease in accumulation from west to east each year. This trend has been observed in previous studies (Fischer et al., 1994; Anklin and Stauffer, 1994; Burgess et al., 2010) and reflects the dominant transport of water vapour in the region from west to east. Inter-annual variability in accumulation has also been observed previously (Fischer et al., 1994). Comparison with historical records suggests that the accumulation patterns have increased by 19% above 3000 m over the last 30 years.

A new generation of radar altimeters have made it possible to obtain accurate estimates of snow accumulation from altimetry by combining the radar observations with field measurements of snow/firn density. Comparison with accumulation rates derived from a meteorological model are favourable although our estimates reveal small scale

variability that cannot be predicted by the model. The differences seen between the model and our estimates provides insight on uncertainties in surface mass balance estimates – a difference of approximately 45 km^3 on average from these results if applied to the dry snow area of the GrIS, significant if compared to the estimated total surface mass balance ranging from 170 to 308 km^3 yr^{-1} (Fettweis, 2007). In the context of the ESA CryoSat-2 mission, designed to obtain very accurate measurements of ice sheet elevation change, the methodology we present will prove particularly useful for improving the accuracy of mass balance estimates of the GrIS, since CryoSat-2 may also have the potential to identify subsurface accumulation layers. The methodology presented here shows the potential that current radar altimeters have to yield reliable estimates of snow accumulation over large regions of the ice sheets which are otherwise difficult to monitor.

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Table 1. Mean annual accumulation rates for the years 1998–2003. Values are in m.w.e.

Year	Mean annual accumulation (m.w.e.)	Modelled mean annual accumulation (m.w.e.)	Decrease in accumulation from 2750 m to 3165 m (%)	Linear regression correlation coefficient (r^2)
2003	0.335 ± 0.034	0.455	18.5	0.826
2002	0.333 ± 0.041	0.440	30.2	0.811
2001	0.343 ± 0.094	0.472	55.4	0.969
2000	0.458 ± 0.045	0.461	14.9	0.729
1999	0.271 ± 0.056	0.220	50.1	0.826
1998	0.396 ± 0.076	0.455	36.0	0.795

Table 2. Mean accumulation rates (in m.w.e.) estimated every 25 km.

Elevation (m)	1998–2003 ASIRAS derived	Fischer et al. (1984–1989)	Anklin et al. (1977–1989)	1998–2003 Modeled
2750	0.461 ± 0.108	0.439 ± 0.088	0.437 ± 0.072	0.486 ± 0.071
2810	0.442 ± 0.080	–	–	0.496 ± 0.074
2870	0.418 ± 0.078	0.385 ± 0.065	0.402 ± 0.049	0.483 ± 0.069
2930	0.393 ± 0.083	–	–	0.457 ± 0.062
2990	0.370 ± 0.069	0.344 ± 0.062	0.349 ± 0.050	0.420 ± 0.049
3050	0.342 ± 0.083	–	–	0.399 ± 0.015
3090	0.322 ± 0.059	–	–	0.375 ± 0.042
3110	0.307 ± 0.063	0.250 ± 0.035	0.249 ± 0.022	0.350 ± 0.037
3150	0.296 ± 0.091	0.229 ± 0.039	0.253 ± 0.030	0.324 ± 0.034
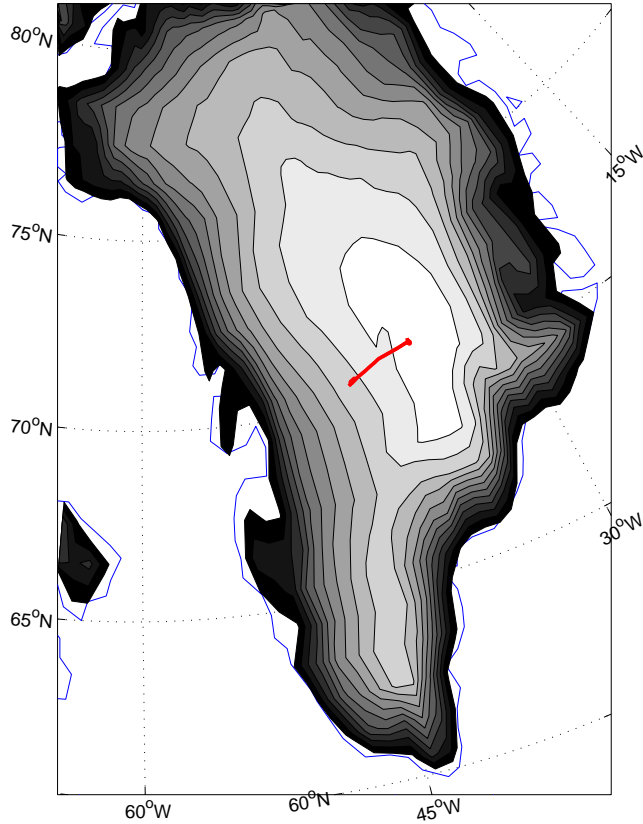
Fig. 1. ASIRAS path over the dry snow region of the GrIS.

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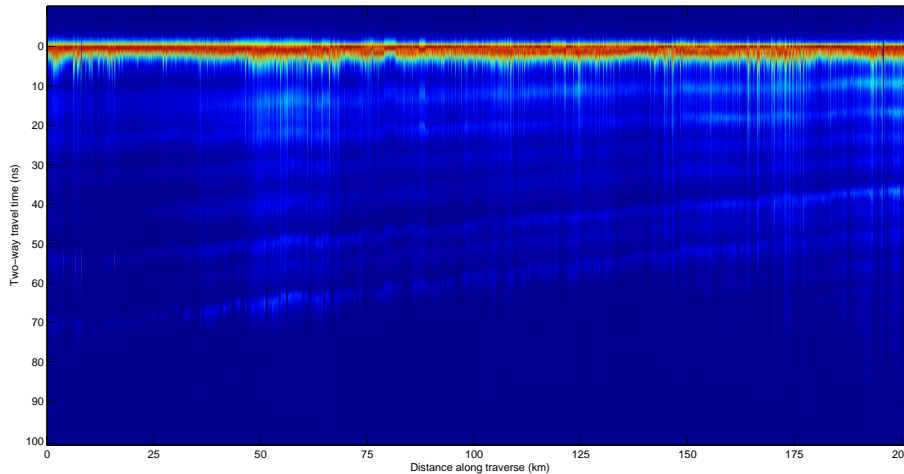
Fig. 2. ASIRAS profiles showing the waveforms recorded over most of the dry region of the ice sheet. The waveforms are centred at the location estimated to be the surface (two way time of 10.5 ns). In the dry zone, the reflection from the surface can be clearly identified. Weaker reflections from the snowpack volume are related to layers of accumulating snow.
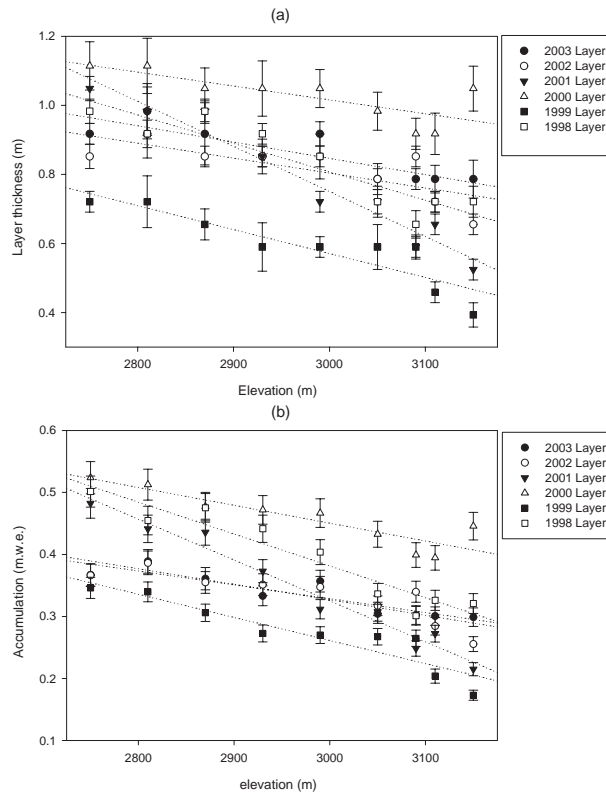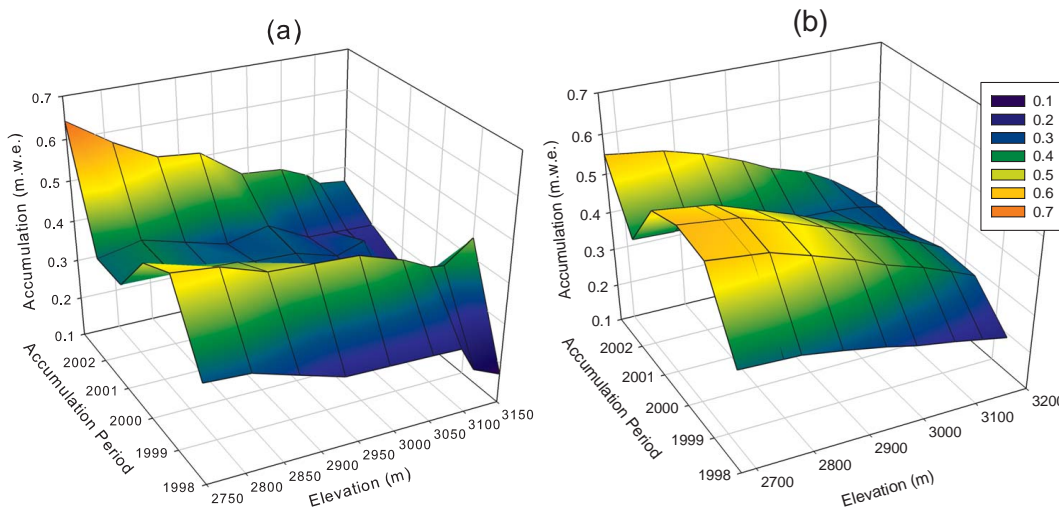
Fig. 3. Seven annual layers are identified corresponding to the annual accumulation for each year from 1998 to 2004. Figure 3a show the thickness for each layer. For scaling purposes, the 2004 accumulation layer is omitted. Figure 3b show the annual snow accumulation estimates obtained from the different layers identified (m.w.e). The estimates were made every 25 km.

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Fig. 4. Observed **(a)** and modelled **(b)** mean annual accumulation rates for 1998–2004.