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Quantifying uncertainty in using multiple datasets to determine spatiotemporal ice mass loss over 101 years at Kårsaglaciären, sub-arctic Sweden

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Abstract

Glacier mass balance and mass balance gradient are fundamentally affected by changes in glacier 3D geometry. Few studies have quantified changing mountain glacier 3D geometry, not least because of a dearth of suitable spatiotemporally-distributed topographical information. Additionally, there can be significant uncertainty in georeferencing of historical data and subsequent calculations of the difference between successive surveys. This study presents multiple 3D glacier reconstructions and the associated mass balance response of Kårsaglaciären, which is a 0.89 ± 0.01 km^2 mountain glacier in sub-arctic Sweden. Reconstructions spanning 101 years were enabled by historical map digitisation and contemporary elevation and thickness surveys. By considering displacements between digitised maps via the identification of common tie-points, uncertainty in both vertical and horizontal planes were estimated. Results demonstrate a long term trend of negative mass balance with an increase in mean elevation, total glacier retreat (1909–2008) of 1311 ± 12 m, and for the period 1926–2010 a volume decrease of 1.0 ± 0.3 x 10^{-3} km^3 yr^{-1}. Synthesising measurements of the glaciers past 3D geometry and ice thickness with theoretically calculated basal stress profiles explains the present thermal regime. The glacier is identified as being disproportionately fast in its rate of mass loss and relative to area, is the fastest retreating glacier in Sweden. Our long-term dataset of glacier 3D geometry changes will be useful for testing models of the evolution of glacier characteristics and behaviour, and ultimately for improving predictions of meltwater production with climate change.

Key words: mountain glacier, mass balance, glacier reconstruction, glacier geometry

Background and rationale

Arctic and sub-arctic glaciers and ice gaps excluding the Greenland ice sheet have a total coverage of 421,791 km^2 (Randolph Glacier Index (RGI) version 5, Pfeffer et al. 2014). Of these glaciers, 87.97 % have an area < 5 km^2 and 78.23 % have an area of < 2 km^2 and 67.70% < 1km^2 (Pfeffer et al. 2015). These small glaciers are important indicators of climate change due to their relatively fast response times when compared to ice sheets (Dyurgerov and Meier 1997, 2000; Bahr et al. 1998, Dyurgerov 2003; Raper and Braithwaite 2006; Haeberli et al. 2007). Furthermore, it is these 'mountain glaciers', i.e. all glaciers globally excluding the Greenland and

Antarctic ice sheets, that contribute most significantly to global sea level rise (Raper and Braithwaite 2006, Gardner et al. 2013). For the period of 2003–2008, glaciers of Greenland separate to the ice sheet were found to have contributed a ~20 % portion of the total mass loss of Greenland (Bolch et al. 2013). Due to the concentration of glaciers globally within the arctic and sub-arctic regions (59.79 % of global glaciers, Pfeffer et al. 2015), the associated fast response times of these glaciers are one reason why the Arctic as a whole is a particularly sensitive region of the World to climate change (Bates et al. 2008).

Global time series of glacier changes are required to reasonably approximate global glacier contribution to sea level (Vaughan et al. 2013). As a means of providing increased spatial coverage information providing for the fifth assessment of the IPCC, the *Randolph Glacier Inventory* (RGI) was developed detailing glacier extent and hypsometry using satellite imagery from 1999-2010 (Pfeffer et al. 2014).

Long term observations of the order of centuries are not available for many of the glaciers in Greenland. However, with increased accessibility of glaciers within the sub-arctic areas of Sweden and Norway over the past century as a result of changes in infrastructure relating to socio-economic pressures and development (Bodin 1993a), many glaciers have been monitored over longer time periods (cf. Holmlund et al. 1996). This provides an opportunity for assessing glacier change over time. Of the glaciers monitored, in-depth analysis of three dimensional glacier change has been focused on a few, larger glaciers including Rabots Glaciär (cf. Brugger et al. 2005) and Storglaciären (e.g. Hock and Holmgren 2005). With the importance of understanding changes in geometry of smaller mountain glaciers, especially those in arctic and sub-arctic environments, the aim of this study is to provide a quantitative analysis of the spatiotemporally distributed mass balance response of a <1 km^2 sub-arctic mountain glacier. In doing so, demonstration of a careful assessment of uncertainty in the workflow for calculating geometrical changes in glaciers is presented. In particular, key obstacles that must be considered when reconstructing a glacier from a variety of data sources are highlighted so that they can be mitigated robustly both in the current and future glacier reconstruction projects.

Reconstructing glacier geometry changes

Studies considering historical glacier change over various periods, some of which also used the glacier areas provided in the RGI, were included in IPCC AR5 (Vaughan et al., 2013) for assessment of glacier contribution to sea level rise. Considering sea level contribution for glaciers globally, excluding those surrounding Greenland and Antarctica, for the longest period reported in IPCC AR5 (Vaughan et al. 2013), a mean annual rate of 0.54 ± 0.07 mm SLE yr^{-1} was reported (the value reported was the average of the combined studies of Marzeion et al. (2012) and Leclercq et al. (2011). The World Glacier Monitoring System (WGMS), RGI and GLIMS (which now includes all WGMS glacier data) databases were the key data sources in the two studies (Arendt et al. 2012 (RGI version 1); GLIMS and NSIDC 2005; Leclercq et al. 2011; Marzeion et al. 2012; Pfeffer et al. 2014 (RGI version 5); WGMS 2014), providing information on glacier length and area changes through time.

Continued data contribution of site-specific glacier measurements including minimum, maximum and median elevation; mean slope and aspect; length; area and hypsometry, and where possible at multiple time periods are essential for ensuring that these databases remain up-to-date. Before it is considered how additional contributions can be made to glacier databases, a review of the data collection approaches taken is pertinent.

Field based observations, encompassing the “traditional” glaciological approach, entail assessment of glacier change from point observations whether for example these be with regard to ablation stake measurements (e.g. Braithwaite 1989, Braithwaite et al. 1998), snow pit analyses, terminus position mapping, point elevations via traditional or modern GPS surveys (e.g. Hodgkins et al. 2007; Zhang et al. 2012) or point thickness measurements through the use of ground penetrating radar(e.g. Rippin et al. 2011; Carrivick et al. 2015). The increasing use and availability of unmanned aerial vehicles (UAVs) (Ryan et al. 2015), surveying equipment of increasing complexity; robotic total stations, terrestrial laser scanners (see Carrivick et al, 2013a, b) as well as the use structure from motion (Westoby et al. 2012) and other photographic based tools all further enhance the toolkit of the field glaciologist. The nature of the point data resultant of these methods is a problem with regard to spatial coverage and ensuring point coverages of appropriate density from which data can be interpolated (Hock and Jensen 1999) effectively and accurately is an issue.

Increased coverage is a key benefit of remote sensing approaches to assess change and involves assessment and integration of satellite images and aerial photographs which can be used to create digital elevation models (DEMs). Subtraction or differencing of successive DEMs enables an (indirect) assessment of glacier mass change by converting volume change to a meltwater equivalent mass (Huss et al. 2013). Differences between the geodetic and traditional (field-measured) glaciological methods have been identified by Hagg et al. (2004) to vary from -0.48 – +0.10 m yr^{-1}. Andreassen (1999) attempted to combine geodetic and glaciological data used to assess Storbreen (Norway) but found that both data sets were prone to large uncertainties, rendering such a comparison void. A minor issue is that the geodetic method does not provide a true mass balance due to an assumption of static ice density (Bamber and Rivera 2007). More problematically, the geodetic method is sensitive to historical map uncertainty - although not eradicating any uncertainties inherent of the data used to create a map, additional user errors can be partially mitigated by producing successive maps using the same exact same approach (Østrem and Haakensen 1999; Koblet et al. 2010). Most importantly, since the geodetic method frequently utilises data from a variety of different sources, varying degrees of uncertainty are introduced in the workflow, particularly with regard to elevation. Consequently, (i) reconstructed glacier surfaces should be separated by a time step sufficiently long enough that observed change is greater than associated uncertainty (Bamber and Rivera 2007), and (ii) the uncertainty associated with each surface should be robustly quantified, and (iii) propagation of the uncertainty through the geodetic workflow should be carefully analysed.

With these factors of data collection in mind, and also considering the implications of the various approaches employed to acquire data detailing glacier geometry and change (e.g. remote sensing versus the traditional glaciological method (cf. Gardner et al. 2013, Kerr 2013)), key outstanding problems relating to uncertainties associated with glacier mass loss assessment include:

1. Volumes are often scaled up based on a few point measurements or the use of centreline analyses (e.g. Shugar et al. 2010) which results in inadequate quantification of spatial variability (Barrand et al. 2010; Berthier et al. 2010). Furthermore, estimates of glacier volume are subject to uncertainties in glacier geometric parameters including area and thickness (Farinotti et al. 2009) – improved knowledge of these uncertainties is therefore important for improving estimates of volume change.

2. Geodetic ice mass loss calculations for individual glaciers are distorted and spatially limited where surface interpolations are based on sparse point networks (Førland and Hanssen-Bauer 2003; Barrand et al. 2010).

3. Knowledge of geometric, topographic and climatic factors of individual glaciers is required to understand and more accurately account for local glacier change (Oerlemans 1987; Granshaw and Fountain 2006; Salinger et al. 2008). Neglecting these factors can result in highly erroneous melt and resultant sea level rise estimates (Barrand et al. 2010). Where observations of glacier characteristics are available, it is particularly important that uncertainty in such assessments is quantified to give realistic quantification of glacier change (cf. Koblet et al. 2013).

4. Poor understanding of the spatiotemporal variability in glacier mass-balance gradients leads to poor melt estimates from modelling routines, as mass-balance sensitivities are inaccurately represented (Raper and Braithwaite 2006).

Study site

Kårsaglaciären (68˚ 21′ N, 18˚ 19′ E), is a small mountain glacier in the Vuoittasrita massif on the border between northern Sweden and Norway (Fig. 1). In 2008, the area of Kårsaglaciären was 0.89 ± 0.01 km^2 (digitized from aerial imagery (Lantmäteriet 2008)) putting it within the same order of magnitude of other glaciers in Scandinavia (~ 0.3 km^2) (RGI (v5), Pfeffer et al. 2014). The presence of Kårsaglaciären has been related to favourable topographical and meteorological conditions; namely that the narrow incised valley catches drifting snow from south westerly winds (Wallén 1949). Following ice penetrating radar surveys of the glacier in 2008/9, it was proposed that Kårsaglaciären exhibited signs of a thermal lag, with its contemporary polythermal state being discordant with its contemporary geometry (Rippin et al. 2011). Climatic conditions at Kårsaglaciären are split between maritime and continental, the maritime conditions often prevailing in the winter months, being replaced by more stable continental conditions during the summer months (Wallén 1948, 1949). Mean July temperatures measured for the period 2007-2011 were $\sim$11°C and mean winter temperatures were $\sim$-10°C with total monthly precipitation measured to be greatest in July and lowest in April.

Kårsaglaciären was selected for this study due to the wealth of available data and information on the glacier throughout the 20th century, as well as its relative accessibility. The amount of data in part owes to the Swedish national mass balance programme initiated in the early 1940s. Data available as a result of past studies on Kårsaglaciären is presented in Table 1. Photographic evidence and visual descriptions evidence glacier terminus advance from 1886 to 1912 with noticeable thickening of the margin (Svenonius 1890-1910;

Sjögren 1909). Since around 1912, the glacier has been in a state of near constant re- treat, with some isolated areas of minor advance (Karlén 1973; Bodin 1993a). Whilst Svenonius (1910) provided a map of the glacier terminus, each successive study; namely Ahlmann and Tryselius (1929), Wallén (1948, 1949); Wallén (1959), Karlén (1973) and Bodin (1993a) provided a topographic map of the glacier outline and of the immediate surrounding area. The surface elevations reported by Wallén (1948, 1949, 1959) were updated via re-georeferencing by Schytt (1963). The last survey of the glacier prior to surveys facilitated for this study starting in 2008 was carried out in 1991.

Methods

Pre-existing data compilation

Historical topographic maps of the Kårsaglaciären (Table 1) are of varying scale and varying quality, where the poorest quality maps had a glacier outline and a stream network but no marked georeferenced points. The historical maps were all derived from summer aerial photography.

We use the term *georectify* here to be explicit about assigning maps from previously digitized aerial photographs (with original geometry therefore not projected in a known coordinate system) to the Swedish national grid coordinate system: RT 90. The 1943 Kårsaglaciären map (Wallén 1948; Schytt 1963) contained the most detail surrounding the glacier. For this reason, the 1943 map was georectified to the lower resolution Lantmäteriet BD6 mountain map (1:100 000), which was projected using the Swedish National grid RT90 gon V. Once the 1943 map georectified in the RT90 gon V coordinate system space, all other maps (as well as the 2008 aerial photograph) were coregistered by matching common features to the 1943 map in turn. All stereo-pair derived images were georectified in the horizontal plane.

All reported elevations on the 1926 – 1991 maps are associated with the sea level at the time of photograph acquisition – no details were available for the specific datums used. It is known that Northern Sweden has an isostatic uplift rate greater than that in the south (Lantmäteriet 2015) however, for the period 1900–2000, this was ~ 1.0 m. Consequently, despite not knowing the precise vertical datum used for map development, this small change in the relative base elevation is not considered in this study.

This georeferencing sequence was implemented to limit georeferencing error to a single quantity where that quantity arose from georeferencing the 1943 1:5000 map to features on the Lantmäteriet 1:100 000 map. The different maps were georeferenced using common features which were limited predominantly to sharp/pronounced inflexions in ridge lines. Both first order and spline transformations were used to assist with coupling reference points between maps, the method resulting in least visible distortion being applied.

Once all maps were georectified in the RT90 system – therefore now being classed as ‘georeferenced’ – they were transformed to the UTM WGS 1984 zone 34N projection to which GPS coordinates from more recent campaigns were easily added (the dGPS data being converted from geographic WGS84 latitude and longitude to the UTM WGS 1984 zone 34N projection).

Contour lines across the glacier and the glacier perimeter were then digitized and converted to points, which were then interpolated to provide continuous representative glacier ice surface elevation grids.

Contemporary data collection and compilation

To extend the data on the geometry of Kårsaglaciären to the present day, an aerial photograph of the glacier was acquired in July 2008 from an altitude of 4800 m a.s.l. (Lantmäteriet pers. comm.). The proglacial region was surveyed using a Leica GPS500 (dGPS) in late August 2007. We use the results of a field survey of the glacier surface elevation carried out in April 2011 to compare 1991 and earlier glacier surface characteristics with the contemporary glacier. The winter survey implemented a Leica GPS500 (dGPS) system mounted on a snowmobile, the survey was carried out in winter as a summer survey would have been difficult on foot resulting in poor data coverage due to large crevassed areas. Under winter conditions the glacier perimeter was virtually impossible (and generally unsafe due to potential avalanching from the adjacent hillslopes and cliffs) to access either by snowmobile or by foot, so glacier perimeter elevations were assumed to match the rock surface at the glacier perimeter and extracted as points from the regional (50 m cell size) DEM, itself gridded using the vertices of contours taken from the Lantmäteriet BD6 mountain map. The dGPS point elevation observations were interpolated to a continuous grid as described below. The points had an internal accuracy of ± 1 m.

To enable comparison between the contemporary and historical glacier datasets, winter snow accumulation was measured by snow probe to assess distributed snowpack thickness (cf. Østrem and Brugman 1991). The thickness values were interpolated across the glacier for each year using a second order polynomial trend interpolation which was then smoothed using a low pass filter. Snow density and snow pack structure was assessed via manually-excavated pits dug during field campaigns for the years 2009, 2010 and 2011, typically one at the lowermost part of the glacier and one in the middle of the glacier, from which a mean snow pack density of 407.13 kg m^{-3} was calculated.

Meteorological records are available 25 km to the east, (and 500 m lower) of Kårsaglaciären at the Abisko Naturvetenskapliga Station (ANS) for the period 1920 – present. These temperatures were lapse rate corrected using observations made from an automatic weather station (AWS) installed at the glacier for the period 2007-2011. The method of temperature correction is described extensively in Williams (2013). For the 1920 – 2010 period, median February and July temperatures of -11˚C and 8˚C were recorded respectively. For the 1920-2010 period, a weak positive increase in median temperature was identified ($p = 0.01$ and adjusted $r^2 = 0.07$).

Meteorological conditions at Kårsaglaciären were recorded for this study using a Campbell Scientific (CR200 logger) automatic weather station (AWS) from spring 2007 to summer 2013. For the 2007–2010 period of this study, median February and July temperatures were -10˚C and 8˚C respectively. Mean total monthly summer (June–August) precipitation for the 2007–2010 period was ~300 mm, with the wettest summer being in 2009 with a total of 360 mm. Precipitation at the AWS was measured using a tipping bucket precipitation collection system which did not allow for assessment of frozen precipitation. Where snow events occurred during the summer, this could potentially have led to local snow accumulation, which following melt may have provided higher precipitation counts than might have been expected. This effect could not be quantified due to the remote nature of the AWS.

The glacier bed – development of an ice free DEM

We use a DEM of bed topography derived from the combination of bed elevations from ice thickness measurement surveys (using GPR) from Bodin (1993a) and Rippin et al. (2011). Ice thickness derived from these surveys were subtracted from measurements of surface elevations at each radar data collection point, with the resultant bed elevation then being interpolated to a regular grid using a kriging interpolation approach.

The Rippin et al. (2011) GPR-derived bed elevation points were estimated to have an internal vertical consistency of 1.4 m based on three data cross over sites (Rippin et al. 2011), the error likely being linked to the method of data collection (movement of the radar and dGPS on the back of a snow mobile). The dGPS data associated with these GPR points had a mean vertical error magnitude of ± 0.07 m. Cross-over analysis was carried out to assess the consistency between the Bodin (1993a) and Rippin et al. (2011) GPR points by comparing measurements within 10 m of one another from which a ± 6.6 m vertical mean cross-over error was identified. The specific methodology describing the combination of these two data sets is detailed in Rippin et al. (2011). Using this bed DEM, we combined it with the regional DEM to provide a seamless regional glacier free topography, encompassing surrounding mountains and land.

Interpolation of glacier surfaces

For the 1926-1991 surfaces, we use regularly spaced vertices derived from along digitized contour lines as input to a kriging algorithm. The same process is repeated for the interpolation of the 2011 dGPS elevation observations. Kriging was chosen as it has been found to work well with data with varying observation density (Bamber et al. 2001) the premise being that an area where the method is being applied is spatially stationary. We use the same method for all datasets to limit the introduction of errors from the use of differing interpolation approaches. The specific variogram models used for the semi-variograms developed for each data set were selected on a case by case basis in the manner outlined in Hock and Jensen (1999). In practice, semivariogram models were chosen with consideration of model statistics that were ranked in order of importance in accordance with ArcGIS tool usage guidelines.

To provide a surface comparable with the 1926–1991 surfaces (all mapped from photographs derived during the summer), we subtracted the interpolated 2011 snowpack thickness grid (described above) from the interpolated 2011 glacier surface, the resultant lowering being an approximation of an increase in elevation as a result of winter snowpack cover. The resultant lowered DEM was then taken as representative of the glacier as of summer 2010. As we approximate the thickness of the snowpack above the glacier only for the winter of 2011, temporal variations in snow accumulation and densification are not considered.

The elevations digitized from the 1926-1991 glacier maps which were interpolated were not altered to account for the presence of a snowpack as there was no known estimate or measurement of distributed snowpack thickness or density.

No known uncertainty values were available for the digitized maps of the glacier for 1926–1991, with a ± 1 m measurement uncertainty being associated with the points used to create the 2010 summer surface. To approximate the uncertainty between interpolated surfaces, we compared the elevation of continuous points along the perimeter of each surface, where we assume ice thickness to be 0 m, and compare this to the ice free DEM. This provided a common reference point against which vertical displacement uncertainty of each surface was approximated (Table 2).

Calculation of temporal change characteristics

Area and hypsometry. Glacier surface area was calculated for each year using digitised glacier outlines as inputs. Original aerial photographs were not available prior to the 2008 study and so the glacier outlines as identified by past cartographers were used. The contemporary area pertains to the 2008 outline as this was the highest resolution (0.5 m) image available closest to the most contemporary stand of the glacier considered in this study. This digitisation was also supported by additional knowledge of the glacier's extent following work carried out at the site during the summer months when snow cover was limited, as well as consulting photographs taken at a similar time to the aerial photograph.

The assessment of glacier area by means of perimeter identification is itself a source of error when considering glacier change (Paul et al. 2013). Due to limited access to original photographs, use of pre-defined glacier extent was all that was possible for the glacier pre-2008. The 2008 perimeter was digitised using aerial photography in tandem with site specific knowledge and planar photographs. We apply a dimensional analysis approach to approximate uncertainty in area assessment based on the width of the glacier perimeter as identified on the original data source maps (and the pixel width of the aerial photograph) to provide upper and lower confidence bounds. Uncertainties relating to the area analysis based on glacier perimeter digitisation uncertainty as a result of image resolution and pixilation (Table 2) accumulated to a < 1 % effect on reported area values. This assessment of perimeter uncertainty translates directly to the assessment of terminus retreat.

Hypsometry for a given year was calculated by parcelling glacier outline polygons into elevation bands using contour lines at 20 m intervals and the glacier perimeter as band boundaries. For each of these bands, elevation-specific areas were then calculated. The hypsometry index was calculated using:

$HI = H_{max} - H_{med} / H_{med} - H_{min}$ (1)

where H_{max} and H_{min} are the maximum and minimum glacier elevations and H_{med} is the elevation of the contour that halves the glacier (Jiskoot et al. 2000, 2009). Resultant values were classified as (Jiskoot et al. 2009):

1. Top heavy (HI < -1.5)

2. Equidimensional (-1.2 < *HI* < 1.2)

3. Bottom heavy (*HI* > 1.5)

Terminus retreat. Various methods exist for the assessment of glacier terminus retreat, often being related to fixed base lines (Lea et al. 2014). In this study, retreat rates were calculated relative to a reference line orientated perpendicular to the west to east flow direction of the lowermost part of Kårsaglaciären (cf. Koblet et al. 2013). The reference line was positioned so as to intersect the most eastern point of the 1909 terminus position, thereby enabling all terminus position retreat values to be calculated relative to the 1909 glacier stance. The most easterly point of the digitised glacier outline (see 'Area and hypsometry') was chosen in line with definitions of flow line locations where the flow line extends from the flow start point to the lowest point in the ice (Giesen and Oerlemans 2010). To provide a retreat value that accounts for spatial variability across the terminus, retreat values were calculated for individual points along the entire length of the terminus defined here as being anywhere within a 300 m buffer of the most eastern point of the glacier flow line and then averaged. There are measurements of terminus retreat made relative to fixed points for the period 1909-1939 reporting retreat of 75.5–131.0m (for different points) however we do not use these fixed points in this study, using a different approach as stated in the methods section 'glacier retreat'. This is done to get a more spatially descriptive estimate. Furthermore, it is unknown between which parts of the glacier terminus and the fixed points the distances were measured which could be a source of large uncertainty. By using the retreat assessment described relative to fully digitised glacier perimeters we attempt to minimise method specific uncertainties.

There are no reported uncertainty values available for the termini derived from the 1926–1991 maps. We provide errors on the assessment of retreat by propagating uncertainties based on values derived from agreement between maps on common tie points used during the georectification process as well as the uncertainty in the terminus position based on the perimeter digitisation (Table 2).

Surface elevation, ice thickness and volume change. Surface elevation change was calculated via the subtraction of each glacier surface DEM from the DEM of the prior time step (e.g. 1926 DEM–1943 DEM). Uncertainty in the vertical plane of each of the DEMs (1926-2010) is based on the perimeter agreement between the elevation at the edge of each DEM and the elevation of the regional ice free DEM, thus providing a relative measure of vertical agreement between all surface DEMs (Table 2). The vertical perimeter agreement of each surface is then propagated using standard quadrature to provide an uncertainty in overall elevation change.

Ice thickness was calculated by subtracting the glacier free DEM from each glacier surface DEM. The cell by cell difference between a glacier surface and the bed DEM provided a distributed ice thickness surface for each year. The differences between the two surface types resulted in some values indicative of negative depths in isolated regions at the margins. These uncertainties were due to combining data sets of different resolution (1:5000 to 1:100 000) and without any means to assess what the values should be, these values were simply changed to zero thickness to avoid further calculation problems. Thickness uncertainty (σ_{thick}) was calculated by propagating measurement uncertainties associated with both the surface elevation of a given time step and the ice free DEM.

The gridded thickness surface for each interval was then used to calculate volume. Volume uncertainty values (σ_{vol}) (Table 2) for the calculated volume for each year was based on the propagation of uncertainty in thickness using the equation:

$$\sigma_{vol} = area \cdot n \cdot \sigma_{thick} \quad (2)$$

where n is the number of cells in a given thickness surface. We assumed a fixed area when calculating volume uncertainty as area uncertainty was found to be <1% of the total for each time step.

Slope. Surface slope is calculated for each glacier surface DEM by considering the maximum rate of change in a 3x3 kernel neighbourhood for each DEM grid cell (Burrough and McDonnell 1998). For each surface, slope is calculated based on the surface interpolation from digitized and observed surface elevation points which are assumed to be internally consistent. Slope calculations are dependent on the accuracy of the DEMs from which they are derived. We consider uncertainties in the DEMs but do not consider associated slope uncertainty (cf. Koblet et al. 2013).

Equilibrium Line Altitude (ELA). The ELA provides a convenient measure of glacier response to changing climate. It can be estimated from various topographic measurements (Osmaston 2005) and un-certainty can be mitigated partly by applying multiple methods (Davies et al. 2012; Carrivick et al. 2015). However, the application of the *Accumulation Area Ratio* (AAR) method is reasonable only where there is mass balance information available. *Balance Ratio* (BR) values borrowed from a nearby glacier will not necessarily be representative because of inter-catchment variability in glacier responses to climate change (e.g. Carrivick and Brewer 2004; Carrivick and Rushmer 2009; Carrivick and Chase 2011). Furthermore, the ELA is a long-term average and so for a single time is only valid under steady state conditions. In this study, median elevations (H_{med}) are considered as an approximation of the ELA following Braithwaite and Raper (2009) who found the balanced-budget ELA to be approximately equal to median glacier altitude. Uncertainties in H_{med} are based on the glacier surface perimeter errors (Table 2).

Basal shear stress (τ_b). Knowledge of the stress exerted by the glacier at the bed can provide information on flow dynamics and has implications for glacier thermal properties (Rippin et al. 2011). τ_b was calculated on a cell by cell basis using the equation:

$$\tau_b = \rho_{ice} \cdot g \cdot h \cdot \sin\alpha \quad (3)$$

where ρ_{ice} is the density of ice (assumed at 900 kg m^{-3}), g is gravitational acceleration (9.81 m s^{-2}), h is ice thickness and α is the surface slope angle (in radians) (Benn and Evans, 1998). Slope was calculated using a moving average equal to twice the glacier mean thickness to smooth out effects of longitudinal and lateral variations (Raymond 1980; Thorp 1991). We calculate τ_b uncertainty (σ_{τ_b}) by propagating thickness uncertainty using the equation:

$$\sigma_{\tau_b} = \rho_{ice} \cdot \mathrm{g} \cdot \sigma_{thick} \cdot \sin\alpha \quad (4)$$

Mass balance (MB). MB was calculated following Cuffey and Paterson (2010) by firstly accounting for the net balance (b_n):

$$b_n = t_2 - t_1 \qquad (5)$$

where t_1 and t_2 represent successive mass balance minimums. The glacier annual balance (B_n) was calculated by the integration of b_n over the glacier surface (A):

$$B_n = \int b_n dA \qquad (6)$$

Specific net balance was then calculated using:

$$\overline{b_n} = \frac{B_n}{A} \quad (7)$$

where b_n was calculated for a surface where data on surface change is only available over large time steps, the value calculated for b_n (equation 5) was then divided by the number of years in the interval between t_1 and t_2. Changes in mass were converted to m w.e. (melt water equivalent) using an assumed ice density of 900 kg m^{-3} (Braithwaite 2002) – we assume there to be no spatiotemporal variability in ice density (no data was available to assess this). Mass balance changes (m w.e) were generalised for elevations on a 1 m interval, and where multiple b_n values exist for a single elevation, a mean value was calculated. Uncertainty values are propagated using standard quadrature based on perimeter uncertainty between each glacier surface and the ice free DEM. MB values are reported as a mean of the values within the area of the glacier as represented at position t_2 (which will be smaller than t_1 where the glacier is reducing in size) which consequently omits losses outside of the area of t_2. This may underestimate total mass balance change compared to considering the extent of the glacier at position t_1. To assess change in terms of area, percentage coverage of both mass loss and gain are also reported.

Results

Kårsaglaciären has been in a state of retreat throughout the 1909–2008 period. The area of the glacier (Fig. 2a and Fig. 4a) reduced a total of 1.69 ± 0.01 km^2 from 2.58 ± 0.03 km^2 in 1926 to 0.89 ± 0.01 km^2 in 2008 (Table 3). Over the same period the glacier retreated (Fig. 2b) by approximately 1.3 ± 0.01 km. Mean annual retreat rates (Table 4) across the glacier terminus were calculated as being greatest between 1943–1959 at 30.3 ± 0.9 m yr^{-1}, being smallest at 8.8 ± 0.8 m yr^{-1} between 1926 and 1943.

Elevation change (Fig. 3) has been predominantly negative, expressed through thinning that has been most pronounced along the glacier centreline, especially in the lowermost part of the glacier (Fig. 4c). Relative to the 2010 glacier extent, the glacier thinned at a mean rate of 0.26 ± 0.1 m yr^{-1}. From 1926–2010, median glacier surface elevation increased from 1170 ± 8.0 m a.s.l. in 1926 to 1236 ± 1.0 m a.s.l. in 2010 (Fig. 2c).

To quantify hypsometry changes through time (Fig. 4b), Hypsometry Index (HI) values are calculated giving values of -1.04, -1.04, 1.17, 1.14, 1.24 and 1.06 for the years 1926, 1943, 1959, 1978, 1991 and 2010, respectively, showing a positive trend ($p = 0.0483$ and adjusted $r^2 = 0.6$). Through characterisation of the glacier by using these *HI* values (Jiskoot et al. 2009), Kårsaglaciären is classed as equidimensional for all years, being more top heavy in 1926 and 1943, relatively equidimensional for 1959, 1978 and 2010 and slightly more top heavy in 1991. There is a noticeable increase in the proportion of elevation >1300 m a.s.l. for 1991 (36%) and 2010 (40%) compared to all other years (19-25%) which shows a major shift in hypsometric distribution (Fig. 4b) that is not identifiable using the HI values alone. Using H_{med} as a pseudo-ELA proxy, the median elevation of the glacier in 2010 (1236 ± 1.0 m a.s.l.) was much greater than in 1926 (1170 ± 8.0 m a.s.l.) indicative of more areal coverage at higher elevations.

Median surface slope (Fig. 2d) increased significantly ($p = 0.0297$, adjusted $r^2 = 0.7$) from 1926–2010 from 14.2◦ to 19.1◦. Reductions in slope also occurred for 1926–43 (-0.9◦) and 1978–91 (-0.4◦). Steepest slopes were continually found to be at the transition between the south-west and central portions of the glacier (Fig. 5). This area is representative of a steep ramp between the uppermost lowermost portions of the glacier. Central and eastern portions of the glacier have much more gradual slopes of between 0 and 20◦. Considering the centreline alone, slope angle increased linearly ($r^2 = 0.998$, $p = 4.2 \times 10^{-6}$) (Figs. 2d).

Mass balance difference surfaces acquired from the assessment of surface elevation change between images were used to calculate changes in glacier mass balance for the different mapping intervals (Fig. 4d). For the majority of the period of interest, the glacier was in a state of negative balance. There is indication of slight positive balance for the period 1978-1991 which is supported by the elevation change plots showing some mass gain around the centre of the glacier body. By far the most negative period identified was 1926–1943, followed closely in magnitude by the period 1959–1978 (Table 4).

As the glacier has retreated and lowered in elevation, there have been large changes in thickness (Table 3 and Figs. 2e and 2f) from a maximum of 142 ± 11 m in 1926 to 56 ± 7 m in 2010. Rates of maximum thickness change have not been constant, being fastest through 1926-1943 at 1.6 ± 0.7 m yr^{-1}, followed by the period 1991-2010 with a rate of change in maximum thickness of 1.3 ± 0.5 m yr^{-1} (Table 4). Median thickness in 2010 (13 ± 7 m) was almost half of that in 1926 (34 ± 11 m). These changes in thickness resulted in large changes in volume (Fig. 4h) from 100.78 ± 0.03 – 13.28 ± 0.01 $km^3 \times 10^{-3}$ (1926 and 2010 respectively) (Table 3). Total ice volume change was 87.5 ± 0.03 $\times 10^{-3}$ km^3 for the 1926–2010 period.

Stress Calculation of τ_b exerted by the glacier over time is indicative of the effect that changing morphology has had on glacier dynamics. Mean (Fig. 4h) reduced between 1926 and 2010 but not in a linear fashion, rather with departures from lower and higher stress values between years. Maximum stress increased more linearly from 1978–2010. For all years, maximum τ_b (Fig. 6) generally occurred between 500 and 700 m from the back of the glacier. There is a secondary peak in τ_b at approximately 1100 m clearly identifiable for 1959-1991. The 1943, 1978 and 1991 stress profiles show much smoother profiles than 1926, 1959 and 2010. τ_b is calculated as having reduced from a maximum of 405 ± 19 kPa in 1926 to 176 ± 13 kPa in 1943, then increasing to 253 ± 25 kPa in 1959 before reducing to 132 ± 14 kPa in 1978. There was then a gradual increase to 169 ± 15 kPa in 2010).

Discussion

Local geometric changes

The 1926–2010 period of investigation is supported as being one dominated by ice mass loss and glacier terminus retreat coupled with an increase in local temperature based on ANS records. The determination of some elevation increases and thus of suggested disparate mass gain are not unexpected and this is supported by considering percentage regions of mass loss and gain rather than using a mean value for the glacier alone. However, considering the spatial element of the analysis presented in terms of total gains and losses, this general trend has not always applied to all regions of the glacier and indeed some periods have shown mean positive mass balance change. This in part is due to the method of spatial mass balance change presented, many parts of the glacier have disappeared between time steps and we present mass balance change for ice still present at the end of a time frame, thus omitting change outside of this area, likely resulting in an underestimate of mass loss. This is not to say that the glacier has not experienced periods of positive mass balance, as for example observed for the years 1989–1990, 1990–1991 and 1991–1992 with net balance values of 0.18, 0.07 and 0.87 m w.e. respectively (with data collected using a traditional stake network) (Bodin 1993a). This same period however falls within the 1991–2010 time step which on average was found to be a period of negative mass balance.

Greater mass loss in the higher-elevation south-westerly part of the glacier than in the central and attitudinally lower part of the glacier is at first glance unusual. However, it can be explained by the exposure to strong and persistent south-westerly winds. Snow depth measurements carried out by this study in the field identified thicker snow in the central region than at higher elevations, and thus we agree with *"sein Licht an beiden Endedn brennt"* (Ahlmann and Tryselius 1929, pp 14) which translates as his light burns at both ends, as recognising wind scour of mass at high elevation on Kårsaglaciären and preferential snow deposition in the topographic lee of the steep slope in the middle part of the glacier.

The steepening slope along the length of the glacier through time, and the coinciding shift in the area-altitude distribution to progressively higher elevations, both have potential to affect the surface energy balance of the glacier. In particular, (i) the intensity of incoming shortwave radiation received at a surface is fundamentally controlled by the slope of that surface. Therefore the availability of such spatiotemporally detailed geometric information as derived by this study is useful for glacier surface energy balance modelling studies, especially when trying to quantify glacier volume changes over longer periods of time and when considering feedbacks between surface topography and surface energy balance (Giesen and Oerlemans 2010; Oerlemans 2010; Carturan et al. 2013).

The polythermal state of Kårsaglaciären (Rippin et al. 2011) is thought to have most likely developed under thicker ice conditions, enabling greater strain-related heating as well as greater insulation against the penetration of colder winter temperatures (Murray et al. 2000; Rippin et al. 2011). In this study, the reconstructions of ice thickness identify thicker ice in the 1926 glacier than at present, and this study also calculates basal stress up to a maximum of 405 ± 19 kpa in 1926 compared to a maximum of 169 ± 15 kPa in

2010. Additionally, the locations of the reconstructed basal stress maxima occur exactly in the region of the glacier that was identified by Rippin et al. (2011) as having greatest radar scatter and thus reasoned to be composed of wetter ice. In brief then, the reconstructions of this study support the hypothesis that the contemporary thermal structure of Kårsaglaciären is an inherited state and not a function of current glacier character or behaviour.

Regional context

The observed surface lowering (based on median thickness) of Kårsaglaciären for the period 1926 to 2010 at -0.8 ± 0.1 m w.e. yr^{-1} is of a similar magnitude to that calculated for nearby Rabots Glaciar for the period of 1910 to 2003 at -0.38 m w.e. yr^{-1} (Brugger et al. 2005). An equivalent surface lowering value of -0.35 m w.e. yr^{-1} was calculated for Storglaciären for the period 1910 to 2001 (Brugger et al. 2005). However, the spatial pattern of change is somewhat different. The general trend, as expected considering elevation change, has been of thickness reduction, but nevertheless with the thickest ice consistently being focused along the centreline of the glacier. This pattern of generally continued thickness reduction is the same as has been identified for nearby Rabots Glaciär (Brugger et al. 2005). In contrast however, and as described above in the results section, Kårsaglaciären has experienced thinning at both higher and lower elevations since the 1990s. Today the greatest ice thickness is north of the glacier centre. Rabots Glaciär and Storglaciären make for fair comparisons as although being larger than Kårsaglaciären, all three glaciers are located near to each other and all share polythermal structures (Brugger et al. 2005). Total ice volume change of 87.5 ± 0.03 x 10^{-3} km^3 for the 1926 to 2010 period can be compared to that of Rabots Glaciär which for the 1910 to 2003 period is reported to have been 153.2 x 10^{-3} km^3 (Brugger et al. 2005). When these changes in volume are converted to annual rates of volume loss, Kårsaglaciären = 1.04 ± 0.0003 x 10-3 km^3 yr^{-1} and Rabots Glaciär = 1.65 x 10^{-3} km^3 yr^{-1}. However, since Kårsaglaciären and Rabots Glaciär have contemporary areas of 0.89 ± 0.01 and 3.70 km^2 respectively, then if these annual rate values are normalised according to the contemporary area ratios of Kårsaglaciären : Rabots Glaciär (0.89 : 3.70), then a respective annual volume change ratio of 4.3 x 10^{-3} km^3 yr^{-1}: 1.65 x 10^{-3} km^3 yr^{-1} can be determined. These calculations demonstrate that whilst annual mass loss at Kårsaglaciären may not be spectacular in absolute terms, relatively for its size Kårsaglaciären has lost a disproportionately high mass of ice.

Comparative analysis of the retreat of the terminus position of Kårsaglaciären relative to other Scandinavian glaciers (Fig. 7) identifies Kårsaglaciären as having re- treated faster than other Swedish glaciers, yet being close to the mean retreat rate of glaciers in Norway (means of -516 and -1246 m respectively). This terminus retreat rate is therefore interesting with regard to Kårsaglaciären's geographical position. With Kårsaglaciären being located more westerly and more northerly than virtually all other glaciers in Sweden, this terminus retreat rate could be indicative of the weather at Kårsaglaciären being dominated by maritime conditions. Indeed Callaghan et al. (2010) suggested that Kårsaglaciären was potentially more susceptible to changes in climate than glaciers in other regions of Scandinavia. This maritime climate-sensitivity hypothesis could further be tested via data on solid precipitation in recent decades at the top of the glacier, which is data that unfortunately does not exist to our knowledge. More widely, the disintegration identified for Kårsaglaciären of a single mountain glacier into several distinct lobes of ice, one of which is probably stagnant is not unique. Disintegration of mountain glaciers through terminus retreat and thinning and outline constrictions has also been reported from the Swiss and Italian Alps (Paul 2004; Citterio et al. 2007).

Absolute area losses for Kårsaglaciären of 0.33 (1978 to 1991) and 0.13 (1991 to 2010) km^2 show a rate reduction which may reflect an increasingly important influence of hill shading, and perhaps also avalanching and debris inputs from the hillslopes, as a greater proportion of total glacier area is situated close to these hillslopes (c.f. Carrivick et al. 2015).

Workflow assessment

Despite great care being taken to ensure upmost reliability in our assessments, some uncertainties still remain. With regard to area, the greatest loss was found for the period 1959–1978 with a total reduction of -0.63 km^2. This timing of area loss may actually be related to the definition of the perimeter of the glacier in the 1978 map, because we have noted that a portion of the northern part of the glacier had subsequently became excluded from subsequent studies. Additionally, the outline of the glacier digitized for 1991 uses a different perimeter to that of Bodin (1993a) and results in a smaller calculated area of the glacier: 1.02 km^2 in this study compared to 1.2 km^2 (Bodin 1993a). The smaller area was reasoned as 1978 mapping of the glacier deemed the part of the glacier outline with discrepancy to be representative of perennial snow. However, mass balance calculations for the separate map intervals (Fig. 4d) do not identify a positive period of mass balance during 1991–92 (Bodin 1993a), the high resolution annual signal having been lost through the averaging process when calculating mean area mass balance profiles (Fig. 4d). This workflow assessment, which highlights how a few subjective or expert decisions must be made, illustrates how such an approach consequently impacts the resolution of mass balance reconstructions.

Quantification of the rate of change of various glacier parameters is key when considering glacier-environment response (e.g. Davies et al. 2012), however where rates are not accompanied by values of uncertainty, confidence in the quantification of report changes is reduced. By propagating errors at each stage of this reconstruction based primarily on uncertainties in elevation measurement and digitisation, uncertainty is associated with calculations of rates of change (cf. Koblet et al. 2013). Reported rates of change in this study are generally greater than associated uncertainty which relates to the frequency of aerial photograph acquisition on from which change is assessed (cf. Bamber and Rivera 2007).

The workflow used in this study analysed stress gradients along the centreline to give an approximation of changing τ_b. The centreline rather than a distributed area approach was applied as ice thickness estimation uncertainties were smaller towards the centre of the glacier than at the margins. The decision to only use stress gradient along the centreline is thus cautionary, given that the method of calculating τ_b is sensitive to slope and thickness input values.

Conclusions

This study has provided the quantitative analysis of the spatiotemporally distributed mass balance response of a small sub-arctic mountain glacier. Through careful consideration of error propagation as a result of the compilation of various data sources, we have presented robust metrics allowing quantification of the effects of changing glacier geometry through time in a warming sub-arctic environment. (1) Observations of changes in glacier geometry have identified a large reduction in area, disintegration, extensive retreat, ice thinning, steepening of the glacier profile and a shift in hypsometric distribution, especially since 1991. (2) These observed changes in glacier geometry – particularly thickness and surface slope – are indicative of changes in $\tau_{b,}$ with larger overall τ_b being associated with the glaciers former extent which we propose explains the polythermal thermal regime lag of the glacier as suggested by Rippin et al. (2011). (3) Locally, we find that the thinning of Kårsaglaciären has been at a rate similar to the much larger nearby Rabots Glaciär, and the rate for both glaciers had been constant through time, unlike for nearby Storglaciären. (4) Regionally, Kårsaglaciären has been retreating at a faster rate than other glaciers in Sweden, however at a slower rate compared to sites in Norway. (5) Quantification of the internal accuracy of individual datasets, and of the propagation of this uncertainty when combining different datasets, which regrettably to date is in general lacking from reconstructions of glacier geometry, provides confidence in the assessment of glacier change from multiple data sources and we hope that our approach can be followed by future equivalent studies.

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Tables

Table 1 Data available for Kårsaglaciären

Year	Data	Source
1884	Terminus photograph	Svenonius 1910
1886	Terminus photograph	Svenonius 1910
1886	Map of the Terminus (1: 6000)	Svenonius 1910
1903–08	Various photographs	Sjögren 1909
1908	Various photographs	Svenonius 1910
1909	Terminus position map (1: 5000)	Svenonius 1910
1917	Terminus measurements	Ahlmann and Lindblad 1940
1919	Terminus measurements	Ahlmann and Lindblad 1940
1920	Map of the terminus (1: 5000)	Ahlmann and Tryselius 1929
1924	Terminus measurements	Ahlmann and Lindblad 1940
1925	Map of the terminus (1 : 15 000)	Ahlmann and Lindblad 1940
1926	Glacier map (1: 15 000)	Ahlmann and Tryselius 1929
1928	Terminus position	Ahlmann and Tryselius 1929
1927–32	Terminus measurements	Ahlmann and Lindblad 1940
1936	Terminus photograph	Ahlmann and Lindblad 1940
1939	Terminus position	Ahlmann and Lindblad 1940
1943	Glacier map (1: 20 000)	Wallén 1948
1942–47	MB study	Wallén 1948
1959	Glacier map (unknown)	University of Stockholm, 1984
1961	Glacier map (unknown)	Schytt, 1963
1978	Glacier map (unknown)	University of Stockholm, 1984
1981–82	MB data	Eriksson (unpublished)
1984–85	MB data	Eriksson (unpublished)
1989–91	MB data	Bodin 1993b
1991–92	GPR survey	Bodin 1993a
1991	Glacier map (unknown)	Bodin 1993a
2008	Aerial photograph	Lantmäteriet (2008)

Table 2 Uncertainty associated with glacier digitisation and associated geometric characteristics

Year	Perimeter (m)	Digitisation (m)	Area (km^2)	Volume ($km^3 \times 10^{-3}$)	Thickness (m)	ELA (H_{med}) (m)	GCP (m)	Stress (kPa)
1909	-	± 5	-	-	-	-	± 10.0	-
1926	± 8.2	± 4	± 0.03	± 0.03	± 10.5	± 8.2	± 10.0	± 19.4
1943	± 0.8	± 7	± 0.04	± 0.01	± 6.6	± 0.8	± 0.0	± 12.8
1959	± 10.9	± 8	± 0.05	± 0.03	± 12.7	± 10.9	± 10.0	± 24.6
1978	± 1.0	± 9	± 0.04	± 0.01	± 6.7	± 1.0	± 10.0	± 14.0
1991	± 1.0	± 5	± 0.02	± 0.01	± 6.7	± 1.0	± 5.0	± 14.6
2010	± 1.0	± 1	± 0.01*	± 0.01	± 6.7	± 1.0	± 5.0	± 15.2

*based on the 2008 perimeter

Table 3 General characteristics of the glacier over the 1926 – 2010 period

Year	**Area** (km^2)	**Median elevation** (m a.s.l.)	**Max. thickness** (m)	**Volume** (km^3 x 10^{-3})	**Max. stress** (kPa)	**Median slope** (degrees)
1926	2.58 ± 0.03	1170.0 ± 8.0	142.0 ± 11.0	100.78 ± 0.03	405.0 ± 19.0	14
1943	2.07 ± 0.04	1170.3 ± 0.8	115.0 ± 7.0	58.37 ± 0.01	176.0 ± 13.0	15
1959	1.98 ± 0.05	1198.0 ± 11.0	111.0 ± 13.0	58.15 ± 0.03	253.0 ± 25.0	15
1978	1.35 ± 0.04	1221.0 ± 1.0	92.0 ± 7.0	26.41 ± 0.01	132.0 ± 14.0	16
1991	1.02 ± 0.02	1233.0 ± 1.0	80.0 ± 7.0	25.67 ± 0.01	159.0 ± 15.0	15
2010	0.89 ± 0.01*	1236.0 ± 1.0	56.0 ± 7.0	13.28 ± 0.01	169.0 ± 15.0	19

*based on the 2008 perimeter

Table 4 Annual rates of change between mapped glacier extents

Period	**Area** (km^2 yr^{-1})	**Max. thickness** (m yr^{-1})	**Volume** (km^3 x 10^{-3} yr^{-1})	**Retreat** (m yr^{-1})	**Mass Balance** (m w.e yr^{-1})
1909 - 1926	-	-	-	2.1 ± 0.9	-
1926 - 1943	0.030 ± 0.003	1.6 ± 0.7	2.493 ± 0.002	8.8 ± 0.8	-1.11 ± 0.4
1943 - 1959	0.006 ± 0.004	0.3 ± 0.1	0.014 ± 0.002	30.3 ± 0.9	-0.12 ± 0.6
1959 - 1978	0.033 ± 0.003	1.0 ± 0.8	1.671 ± 0.001	11.2 ± 1.0	-0.99 ± 0.5
1978 - 1991	0.025 ± 0.003	0.9 ± 0.7	0.057 ± 0.001	13.0 ± 1.2	0.19 ± 0.1
1991 - 2010	0.008 ± 0.001	1.3 ± 0.5	0.700 ± 0.0005	15.3 ± 0.5	-0.28 ± 0.1
1909 - 2010	0.02 ± 0.01*	1.0 ± 0.1	1.042 ± 0.0003	13.2 ± 0.1	-0.4 ± 0.1

*based on the 2008 perimeter

Figures

Figure 1: Kårsaglaciären, northern Sweden (68° 21’ N, 18° 19’ E).

Figure 2: Summary plots showing change in area (A); terminus retreat (B); maximum (solid), mean (dot-dash) and minimum (dashed) elevation (C); median slope (D); mean thickness (E); maximum thickness (F); volume (G); mean stress (H). Uncertainties are applied to data according to the values presented in Table 3.

Figure 3: Elevation change per map interval. Reliability plots are displayed for each interval difference map as calculated based on the uncertainties presented in Table 3.

Figure 4: Glacier area in 1926 (with inclusion of an approximation of the side glacier not quantified in this study) and 2008 the centreline is identified by the dashed line (A); Hypsometry curves for 1926-2010 (B); Long profile along the centreline of the glacier 1926-2010 (C); Annual mass balance curves for the glacier per map interval (D).

Figure 5: Slope maps calculated from elevation profiles 1926-2010.

Figure 6: Stress profiles taken along the centreline of the glacier (see Fig. 4a).

Figure 7: Relative terminus retreat totals for glaciers in the political zones of Sweden and Norway. Only glaciers with records >90 years are presented for brevity and clarity (WGMS 2014).