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

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## 10 **Abstract**

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

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

## 28 **Background and rationale**

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

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

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

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

## 58 **Reconstructing glacier geometry changes**

59 Studies considering historical glacier change over various periods, some of which also used the glacier areas  
60 provided in the RGI, were included in IPCC AR5 (Vaughan et al., 2013) for assessment of glacier contribution to  
61 sea level rise. Considering sea level contribution for glaciers globally, excluding those surrounding Greenland  
62 and Antarctica, for the longest period reported in IPCC AR5 (Vaughan et al. 2013), a mean annual rate of  $0.54 \pm$   
63  $0.07$  mm SLE yr<sup>-1</sup> was reported (the value reported was the average of the combined studies of Marzeion et al.  
64 (2012) and Leclercq et al. (2011). The World Glacier Monitoring System (WGMS), RGI and GLIMS (which now  
65 includes all WGMS glacier data) databases were the key data sources in the two studies (Arendt et al. 2012  
66 (RGI version 1); GLIMS and NSIDC 2005; Leclercq et al. 2011; Marzeion et al. 2012; Pfeffer et al. 2014 (RGI  
67 version 5); WGMS 2014), providing information on glacier length and area changes through time.

68 Continued data contribution of site-specific glacier measurements including minimum, maximum and median  
69 elevation; mean slope and aspect; length; area and hypsometry, and where possible at multiple time periods

70 are essential for ensuring that these databases remain up-to-date. Before it is considered how additional  
71 contributions can be made to glacier databases, a review of the data collection approaches taken is pertinent.

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

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

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

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

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

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

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

#### 121 **Study site**

122 Kårsagläciären (68° 21' N, 18° 19' E), is a small mountain glacier in the Vuottasrita massif on the border  
123 between northern Sweden and Norway (Fig. 1). In 2008, the area of Kårsagläciären was  $0.89 \pm 0.01 \text{ km}^2$   
124 (digitized from aerial imagery (Lantmäteriet 2008)) putting it within the same order of magnitude of other  
125 glaciers in Scandinavia ( $\sim 0.3 \text{ km}^2$ ) (RGI (v5), Pfeffer et al. 2014). The presence of Kårsagläciären has been  
126 related to favourable topographical and meteorological conditions; namely that the narrow incised valley  
127 catches drifting snow from south westerly winds (Wallén 1949). Following ice penetrating radar surveys of the  
128 glacier in 2008/9, it was proposed that Kårsagläciären exhibited signs of a thermal lag, with its contemporary  
129 polythermal state being discordant with its contemporary geometry (Rippin et al. 2011). Climatic conditions at  
130 Kårsagläciären are split between maritime and continental, the maritime conditions often prevailing in the  
131 winter months, being replaced by more stable continental conditions during the summer months (Wallén  
132 1948, 1949). Mean July temperatures measured for the period 2007-2011 were  $\sim 11^\circ\text{C}$  and mean winter  
133 temperatures were  $\sim -10^\circ\text{C}$  with total monthly precipitation measured to be greatest in July and lowest in  
134 April.

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

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

## 147 **Methods**

### 148 *Pre-existing data compilation*

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

152 We use the term *georectify* here to be explicit about assigning maps from previously digitized aerial  
153 photographs (with original geometry therefore not projected in a known coordinate system) to the Swedish  
154 national grid coordinate system: RT 90. The 1943 Kårsaglaciären map (Wallén 1948; Schytt 1963) contained the  
155 most detail surrounding the glacier. For this reason, the 1943 map was georectified to the lower resolution  
156 Lantmäteriet BD6 mountain map (1:100 000), which was projected using the Swedish National grid RT90 gon  
157 V. Once the 1943 map georectified in the RT90 gon V coordinate system space, all other maps (as well as the  
158 2008 aerial photograph) were coregistered by matching common features to the 1943 map in turn. All stereo-  
159 pair derived images were georectified in the horizontal plane.

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

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

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

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

#### 176 **Contemporary data collection and compilation**

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

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

197 Meteorological records are available 25 km to the east, (and 500 m lower) of Kårsaglaciären at the Abisko  
198 Naturvetenskapliga Station (ANS) for the period 1920 – present. These temperatures were lapse rate corrected  
199 using observations made from an automatic weather station (AWS) installed at the glacier for the period 2007-  
200 2011. The method of temperature correction is described extensively in Williams (2013). For the 1920 – 2010  
201 period, median February and July temperatures of  $-11^{\circ}\text{C}$  and  $8^{\circ}\text{C}$  were recorded respectively. For the 1920-  
202 2010 period, a weak positive increase in median temperature was identified ( $p = 0.01$  and adjusted  $r^2 = 0.07$ ).

203 Meteorological conditions at Kårsaglaciären were recorded for this study using a Campbell Scientific (CR200  
204 logger) automatic weather station (AWS) from spring 2007 to summer 2013. For the 2007–2010 period of this  
205 study, median February and July temperatures were  $-10^{\circ}\text{C}$  and  $8^{\circ}\text{C}$  respectively. Mean total monthly summer  
206 (June–August) precipitation for the 2007–2010 period was  $\sim 300$  mm, with the wettest summer being in 2009  
207 with a total of 360 mm. Precipitation at the AWS was measured using a tipping bucket precipitation collection  
208 system which did not allow for assessment of frozen precipitation. Where snow events occurred during the  
209 summer, this could potentially have led to local snow accumulation, which following melt may have provided

210 higher precipitation counts than might have been expected. This effect could not be quantified due to the  
211 remote nature of the AWS.

#### 212 *The glacier bed – development of an ice free DEM*

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

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

#### 227 *Interpolation of glacier surfaces*

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

237 To provide a surface comparable with the 1926–1991 surfaces (all mapped from photographs derived during  
238 the summer), we subtracted the interpolated 2011 snowpack thickness grid (described above) from the  
239 interpolated 2011 glacier surface, the resultant lowering being an approximation of an increase in elevation as  
240 a result of winter snowpack cover. The resultant lowered DEM was then taken as representative of the glacier  
241 as of summer 2010. As we approximate the thickness of the snowpack above the glacier only for the winter of  
242 2011, temporal variations in snow accumulation and densification are not considered.

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

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

### 252 *Calculation of temporal change characteristics*

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

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

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

$$273 \quad HI = H_{\max} - H_{\text{med}} / H_{\text{med}} - H_{\min} \quad (1)$$

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

276 1. Top heavy ( $HI < -1.5$ )

277 2. Equidimensional ( $-1.2 < HI < 1.2$ )

278 3. Bottom heavy ( $HI > 1.5$ )

279 *Terminus retreat.* Various methods exist for the assessment of glacier terminus retreat, often being related to  
280 fixed base lines (Lea et al. 2014). In this study, retreat rates were calculated relative to a reference line  
281 orientated perpendicular to the west to east flow direction of the lowermost part of Kårsaglaciären (cf. Koblet  
282 et al. 2013). The reference line was positioned so as to intersect the most eastern point of the 1909 terminus  
283 position, thereby enabling all terminus position retreat values to be calculated relative to the 1909 glacier  
284 stance. The most easterly point of the digitised glacier outline (see 'Area and hypsometry') was chosen in line  
285 with definitions of flow line locations where the flow line extends from the flow start point to the lowest point  
286 in the ice (Giesen and Oerlemans 2010). To provide a retreat value that accounts for spatial variability across  
287 the terminus, retreat values were calculated for individual points along the entire length of the terminus  
288 defined here as being anywhere within a 300 m buffer of the most eastern point of the glacier flow line and  
289 then averaged. There are measurements of terminus retreat made relative to fixed points for the period 1909-  
290 1939 reporting retreat of 75.5–131.0m (for different points) however we do not use these fixed points in this  
291 study, using a different approach as stated in the methods section 'glacier retreat'. This is done to get a more  
292 spatially descriptive estimate. Furthermore, it is unknown between which parts of the glacier terminus and the  
293 fixed points the distances were measured which could be a source of large uncertainty. By using the retreat  
294 assessment described relative to fully digitised glacier perimeters we attempt to minimise method specific  
295 uncertainties.

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

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

307 Ice thickness was calculated by subtracting the glacier free DEM from each glacier surface DEM. The cell by cell  
308 difference between a glacier surface and the bed DEM provided a distributed ice thickness surface for each  
309 year. The differences between the two surface types resulted in some values indicative of negative depths in  
310 isolated regions at the margins. These uncertainties were due to combining data sets of different resolution  
311 (1:5000 to 1:100 000) and without any means to assess what the values should be, these values were simply  
312 changed to zero thickness to avoid further calculation problems. Thickness uncertainty ( $\sigma_{thick}$ ) was calculated

313 by propagating measurement uncertainties associated with both the surface elevation of a given time step and  
314 the ice free DEM.

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

$$318 \quad \sigma_{vol} = area \cdot n \cdot \sigma_{thick} \quad (2)$$

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

321 *Slope*. Surface slope is calculated for each glacier surface DEM by considering the maximum rate of change in a  
322 3x3 kernel neighbourhood for each DEM grid cell (Burrough and McDonnell 1998). For each surface, slope is  
323 calculated based on the surface interpolation from digitized and observed surface elevation points which are  
324 assumed to be internally consistent. Slope calculations are dependent on the accuracy of the DEMs from which  
325 they are derived. We consider uncertainties in the DEMs but do not consider associated slope uncertainty (cf.  
326 Koblet et al. 2013).

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

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

$$341 \quad \tau_b = \rho_{ice} \cdot g \cdot h \cdot \sin \alpha \quad (3)$$

342 where  $\rho_{ice}$  is the density of ice (assumed at 900 kg m<sup>-3</sup>),  $g$  is gravitational acceleration (9.81 m s<sup>-2</sup>),  $h$  is ice  
343 thickness and  $\alpha$  is the surface slope angle (in radians) (Benn and Evans, 1998). Slope was calculated using a  
344 moving average equal to twice the glacier mean thickness to smooth out effects of longitudinal and lateral  
345 variations (Raymond 1980; Thorp 1991). We calculate  $\tau_b$  uncertainty ( $\sigma_{\tau_b}$ ) by propagating thickness uncertainty  
346 using the equation:

347 
$$\sigma_{\tau_b} = \rho_{ice} \cdot g \cdot \sigma_{thick} \cdot \sin \alpha \quad (4)$$

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

350 
$$b_n = t_2 - t_1 \quad (5)$$

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

353 
$$B_n = \int b_n dA \quad (6)$$

354 Specific net balance was then calculated using:

355 
$$\bar{b}_n = \frac{B_n}{A} \quad (7)$$

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

## 368 **Results**

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

374 Elevation change (Fig. 3) has been predominantly negative, expressed through thinning that has been most  
375 pronounced along the glacier centreline, especially in the lowermost part of the glacier (Fig. 4c). Relative to the  
376 2010 glacier extent, the glacier thinned at a mean rate of  $0.26 \pm 0.1 \text{ m yr}^{-1}$ . From 1926–2010, median glacier  
377 surface elevation increased from  $1170 \pm 8.0 \text{ m a.s.l.}$  in 1926 to  $1236 \pm 1.0 \text{ m a.s.l.}$  in 2010 (Fig. 2c).

378 To quantify hypsometry changes through time (Fig. 4b), Hypsometry Index (HI) values are calculated giving  
379 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,

380 respectively, showing a positive trend ( $p = 0.0483$  and adjusted  $r^2 = 0.6$ ). Through characterisation of the  
381 glacier by using these  $HI$  values (Jiskoot et al. 2009), Kårsaglaciären is classed as equidimensional for all years,  
382 being more top heavy in 1926 and 1943, relatively equidimensional for 1959, 1978 and 2010 and slightly more  
383 top heavy in 1991. There is a noticeable increase in the proportion of elevation  $>1300$  m a.s.l. for 1991 (36%)  
384 and 2010 (40%) compared to all other years (19–25%) which shows a major shift in hypsometric distribution  
385 (Fig. 4b) that is not identifiable using the  $HI$  values alone. Using  $H_{med}$  as a pseudo-ELA proxy, the median  
386 elevation of the glacier in 2010 ( $1236 \pm 1.0$  m a.s.l.) was much greater than in 1926 ( $1170 \pm 8.0$  m a.s.l.)  
387 indicative of more areal coverage at higher elevations.

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

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

400 As the glacier has retreated and lowered in elevation, there have been large changes in thickness (Table 3 and  
401 Figs. 2e and 2f) from a maximum of  $142 \pm 11$  m in 1926 to  $56 \pm 7$  m in 2010. Rates of maximum thickness  
402 change have not been constant, being fastest through 1926–1943 at  $1.6 \pm 0.7$  m  $\text{yr}^{-1}$ , followed by the period  
403 1991–2010 with a rate of change in maximum thickness of  $1.3 \pm 0.5$  m  $\text{yr}^{-1}$  (Table 4). Median thickness in 2010  
404 ( $13 \pm 7$  m) was almost half of that in 1926 ( $34 \pm 11$  m). These changes in thickness resulted in large changes in  
405 volume (Fig. 4h) from  $100.78 \pm 0.03 - 13.28 \pm 0.01$  km<sup>3</sup>  $\times 10^{-3}$  (1926 and 2010 respectively) (Table 3). Total ice  
406 volume change was  $87.5 \pm 0.03 \times 10^{-3}$  km<sup>3</sup> for the 1926–2010 period.

407 Stress Calculation of  $\tau_b$  exerted by the glacier over time is indicative of the effect that changing morphology  
408 has had on glacier dynamics. Mean (Fig. 4h) reduced between 1926 and 2010 but not in a linear fashion, rather  
409 with departures from lower and higher stress values between years. Maximum stress increased more linearly  
410 from 1978–2010. For all years, maximum  $\tau_b$  (Fig. 6) generally occurred between 500 and 700 m from the back  
411 of the glacier. There is a secondary peak in  $\tau_b$  at approximately 1100 m clearly identifiable for 1959–1991. The  
412 1943, 1978 and 1991 stress profiles show much smoother profiles than 1926, 1959 and 2010.  $\tau_b$  is calculated as  
413 having reduced from a maximum of  $405 \pm 19$  kPa in 1926 to  $176 \pm 13$  kPa in 1943, then increasing to  $253 \pm 25$   
414 kPa in 1959 before reducing to  $132 \pm 14$  kPa in 1978. There was then a gradual increase to  $169 \pm 15$  kPa in  
415 2010).

416 **Discussion**

417 *Local geometric changes*

418 The 1926–2010 period of investigation is supported as being one dominated by ice mass loss and glacier  
419 terminus retreat coupled with an increase in local temperature based on ANS records. The determination of  
420 some elevation increases and thus of suggested disparate mass gain are not unexpected and this is supported  
421 by considering percentage regions of mass loss and gain rather than using a mean value for the glacier alone.  
422 However, considering the spatial element of the analysis presented in terms of total gains and losses, this  
423 general trend has not always applied to all regions of the glacier and indeed some periods have shown mean  
424 positive mass balance change. This in part is due to the method of spatial mass balance change presented,  
425 many parts of the glacier have disappeared between time steps and we present mass balance change for ice  
426 still present at the end of a time frame, thus omitting change outside of this area, likely resulting in an  
427 underestimate of mass loss. This is not to say that the glacier has not experienced periods of positive mass  
428 balance, as for example observed for the years 1989–1990, 1990–1991 and 1991–1992 with net balance values  
429 of 0.18, 0.07 and 0.87 m w.e. respectively (with data collected using a traditional stake network) (Bodin  
430 1993a). This same period however falls within the 1991–2010 time step which on average was found to be a  
431 period of negative mass balance.

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

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

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

452 2010. Additionally, the locations of the reconstructed basal stress maxima occur exactly in the region of the  
453 glacier that was identified by Rippin et al. (2011) as having greatest radar scatter and thus reasoned to be  
454 composed of wetter ice. In brief then, the reconstructions of this study support the hypothesis that the  
455 contemporary thermal structure of Kårsaglaciären is an inherited state and not a function of current glacier  
456 character or behaviour.

#### 457 *Regional context*

458 The observed surface lowering (based on median thickness) of Kårsaglaciären for the period 1926 to 2010 at -  
459  $0.8 \pm 0.1$  m w.e.  $\text{yr}^{-1}$  is of a similar magnitude to that calculated for nearby Rabots Glaciär for the period of  
460 1910 to 2003 at  $-0.38$  m w.e.  $\text{yr}^{-1}$  (Brugger et al. 2005). An equivalent surface lowering value of  $-0.35$  m w.e.  $\text{yr}^{-1}$   
461 was calculated for Storglaciären for the period 1910 to 2001 (Brugger et al. 2005). However, the spatial  
462 pattern of change is somewhat different. The general trend, as expected considering elevation change, has  
463 been of thickness reduction, but nevertheless with the thickest ice consistently being focused along the  
464 centreline of the glacier. This pattern of generally continued thickness reduction is the same as has been  
465 identified for nearby Rabots Glaciär (Brugger et al. 2005). In contrast however, and as described above in the  
466 results section, Kårsaglaciären has experienced thinning at both higher and lower elevations since the 1990s.  
467 Today the greatest ice thickness is north of the glacier centre. Rabots Glaciär and Storglaciären make for fair  
468 comparisons as although being larger than Kårsaglaciären, all three glaciers are located near to each other and  
469 all share polythermal structures (Brugger et al. 2005). Total ice volume change of  $87.5 \pm 0.03 \times 10^{-3}$   $\text{km}^3$  for the  
470 1926 to 2010 period can be compared to that of Rabots Glaciär which for the 1910 to 2003 period is reported  
471 to have been  $153.2 \times 10^{-3}$   $\text{km}^3$  (Brugger et al. 2005). When these changes in volume are converted to annual  
472 rates of volume loss, Kårsaglaciären =  $1.04 \pm 0.0003 \times 10^{-3}$   $\text{km}^3 \text{yr}^{-1}$  and Rabots Glaciär =  $1.65 \times 10^{-3}$   $\text{km}^3 \text{yr}^{-1}$ .  
473 However, since Kårsaglaciären and Rabots Glaciär have contemporary areas of  $0.89 \pm 0.01$  and  $3.70$   $\text{km}^2$   
474 respectively, then if these annual rate values are normalised according to the contemporary area ratios of  
475 Kårsaglaciären : Rabots Glaciär ( $0.89 : 3.70$ ), then a respective annual volume change ratio of  $4.3 \times 10^{-3}$   $\text{km}^3 \text{yr}^{-1}$   
476  $: 1.65 \times 10^{-3}$   $\text{km}^3 \text{yr}^{-1}$  can be determined. These calculations demonstrate that whilst annual mass loss at  
477 Kårsaglaciären may not be spectacular in absolute terms, relatively for its size Kårsaglaciären has lost a  
478 disproportionately high mass of ice.

479 Comparative analysis of the retreat of the terminus position of Kårsaglaciären relative to other Scandinavian  
480 glaciers (Fig. 7) identifies Kårsaglaciären as having retreated faster than other Swedish glaciers, yet being  
481 close to the mean retreat rate of glaciers in Norway (means of  $-516$  and  $-1246$  m respectively). This terminus  
482 retreat rate is therefore interesting with regard to Kårsaglaciären's geographical position. With Kårsaglaciären  
483 being located more westerly and more northerly than virtually all other glaciers in Sweden, this terminus  
484 retreat rate could be indicative of the weather at Kårsaglaciären being dominated by maritime conditions.  
485 Indeed Callaghan et al. (2010) suggested that Kårsaglaciären was potentially more susceptible to changes in  
486 climate than glaciers in other regions of Scandinavia. This maritime climate-sensitivity hypothesis could further  
487 be tested via data on solid precipitation in recent decades at the top of the glacier, which is data that  
488 unfortunately does not exist to our knowledge. More widely, the disintegration identified for Kårsaglaciären of

489 a single mountain glacier into several distinct lobes of ice, one of which is probably stagnant is not unique.  
490 Disintegration of mountain glaciers through terminus retreat and thinning and outline constrictions has also  
491 been reported from the Swiss and Italian Alps (Paul 2004; Citterio et al. 2007).

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

#### 496 *Workflow assessment*

497 Despite great care being taken to ensure utmost reliability in our assessments, some uncertainties still remain.  
498 With regard to area, the greatest loss was found for the period 1959–1978 with a total reduction of -0.63 km<sup>2</sup>.  
499 This timing of area loss may actually be related to the definition of the perimeter of the glacier in the 1978  
500 map, because we have noted that a portion of the northern part of the glacier had subsequently become  
501 excluded from subsequent studies. Additionally, the outline of the glacier digitized for 1991 uses a different  
502 perimeter to that of Bodin (1993a) and results in a smaller calculated area of the glacier: 1.02 km<sup>2</sup> in this study  
503 compared to 1.2 km<sup>2</sup> (Bodin 1993a). The smaller area was reasoned as 1978 mapping of the glacier deemed  
504 the part of the glacier outline with discrepancy to be representative of perennial snow. However, mass balance  
505 calculations for the separate map intervals (Fig. 4d) do not identify a positive period of mass balance during  
506 1991–92 (Bodin 1993a), the high resolution annual signal having been lost through the averaging process when  
507 calculating mean area mass balance profiles (Fig. 4d). This workflow assessment, which highlights how a few  
508 subjective or expert decisions must be made, illustrates how such an approach consequently impacts the  
509 resolution of mass balance reconstructions.

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

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

#### 522 **Conclusions**

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

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757 Table 1 Data available for Kårsaglaciären

<b>Year</b>	<b>Data</b>	<b>Source</b>
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)

759 Table 2 Uncertainty associated with glacier digitisation and associated geometric characteristics

760

<b>Year</b>	<b>Perimeter (m)</b>	<b>Digitisation (m)</b>	<b>Area (km<sup>2</sup>)</b>	<b>Volume (km<sup>3</sup> x 10<sup>-3</sup>)</b>	<b>Thickness (m)</b>	<b>ELA (H<sub>med</sub>) (m)</b>	<b>GCP (m)</b>	<b>Stress (kPa)</b>
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

761

762

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

<b>Year</b>	<b>Area (km<sup>2</sup>)</b>	<b>Median elevation (m a.s.l.)</b>	<b>Max. thickness (m)</b>	<b>Volume (km<sup>3</sup> x 10<sup>-3</sup>)</b>	<b>Max. stress (kPa)</b>	<b>Median slope (degrees)</b>
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

764

765

766 Table 4 Annual rates of change between mapped glacier extents

<b>Period</b>	<b>Area (km<sup>2</sup> yr<sup>-1</sup>)</b>	<b>Max. thickness (m yr<sup>-1</sup>)</b>	<b>Volume (km<sup>3</sup> x 10<sup>-3</sup> yr<sup>-1</sup>)</b>	<b>Retreat (m yr<sup>-1</sup>)</b>	<b>Mass Balance (m w.e yr<sup>-1</sup>)</b>
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

767

768

769 **Figures**

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

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

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

777 Figure 4: Glacier area in 1926 (with inclusion of an approximation of the side glacier  
778 not quantified in this study) and 2008 the centreline is identified by the dashed line (A); Hypsometry curves for  
779 1926-2010 (B); Long profile along the centreline of the glacier 1926-2010 (C); Annual mass balance curves for  
780 the glacier per map interval (D).

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

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

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