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# CHANGES IN THE GEOMETRY AND VOLUME OF RABOTS GLACIÄR, SWEDEN, 2003–2011: RECENT ACCELERATED VOLUME LOSS LINKED TO MORE NEGATIVE SUMMER BALANCES

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Brugger, K.A. and Pankratz, L., 2014. Changes in the geometry and volume of Rabots glaciär, Sweden, 2003–2011: recent accelerated volume loss linked to more negative summer balances. *Geografiska Annaler: Series A, Physical Geography*, ••, •••••. DOI:10.1111/geoa.12062

**ABSTRACT.** Terminus geometry, ice margins, and surface elevations on Rabots glaciär were measured using differential GPS during summer 2011 and compared with those similarly measured in 2003. Glacier length over the eight years decreased by ~105 m corresponding to  $13 \text{ m a}^{-1}$ , a rate consistent with ice recession over the last several decades. Measured changes in surface elevations show that between 2003 and 2011 the glacier's volume decreased by  $\sim 27.6 \pm 2.6 \times 10^6 \text{ m}^3$ , or  $3.5 \pm 0.3 \times 10^6 \text{ m}^3 \text{ a}^{-1}$ . This compares favorably with an estimate of  $-28.1 \pm 2.6 \times 10^6 \text{ m}^3$  based on a mass-balance approach. The rate of volume loss appears, however, to have significantly increased after 2003, being substantially greater than rates determined for the intervals 1959–80, 1980–89, and 1989–2003. This increase corresponds to a sustained interval of more negative summer balances. Previous work suggests that as of 2003 Rabots glaciär had not yet completed its response to a  $\sim 1^\circ\text{C}$  warming that occurred c. 1900, and thus the current marked increase rate of ice loss might reflect the effect of recent, or accelerated regional warming that occurred during the last decade superimposed on its continued response to that earlier warming.

**Key words:** glacier, climate, mass balance, dGPS, Rabots glaciär, northern Sweden

## Introduction

Given the paucity of meteorological records, glaciers in high, northern latitudes are particularly important indicators of regional climate change (Dowdeswell *et al.* 1997; Dyurgerov and Meier 2000; Lemke *et al.* 2007) and their retreat is recognized as making a significant contribution to current sea-level rise (Meier *et al.* 2007; Radić and Hock 2011; Gardner *et al.* 2011). There is, however, an

insufficient number of long-term, continuous (i.e. annual, or nearly so) measurements of recent glacier change and/or mass balance to fully understand the impact of a changing climate on the Arctic cryosphere; specifically there exists a need for increasing the geographic coverage and extending the time series of such measurements (Dyurgerov and Meier 2000; Lemke *et al.* 2007; Zemp *et al.* 2009). Consequently geodetic measurements of glacier surfaces, typically repeated over multiyear intervals, have become increasingly useful to document global and regional changes in ice volume and mass (e.g. Cogley 2009; Nuth *et al.* 2010). These measurements afford a better assessment of glacier response to climate change than do, for example, records of glacier length variations because of the various factors that can influence the behavior of the glacier's terminus. Geodetic measurements have also been useful for reconstructing annual mass-balance variations (Vincent 2002) and for evaluation of uncertainties associated with the direct, or glaciological method of balance measurement or vice versa (Krimmel 1999; Cogley 2009; Zemp *et al.* 2010). Moreover, even in those regions where long-term records exist (e.g. northern Scandinavia), such measurements could help resolve the ambiguity resulting from disparate glacier behavior within a restricted geographic area (Brugger 2007) in order to better assess and clarify the impact of climate change.

In this paper we present the results of a second differential GPS (dGPS) survey undertaken in 24–29 Jul., 2011 to determine the changes in the geometry and volume of Rabots glaciär that have occurred since the first dGPS survey carried out in 17–21 Jul., 2003 (Brugger *et al.* 2005). In addition to building on a long observational record (going

back slightly more than 100 years), we compare our results with volume changes calculated using a record of annual mass-balance variations for the same period. We then briefly discuss the changes in geometry in the context of the glacier's shorter- and longer-term behavior, and mass balance variations driven by ongoing climate change.

*Physical setting and recent glacier behavior*

Rabots glaciär (67°55'N, 18°30'E) is located in the Kebnekaise Massif in northern Sweden (Fig. 1) and as of 2011 had an area (planar) of about 3.4 km<sup>2</sup>. Like others in northern Sweden, the glacier advanced late in the nineteenth century during the local culmination of the Little Ice Age (LIA) (Karlén 1973; Holmlund 1993), and then began to retreat in response to a regional warming of ~1°C for both annual and summer temperatures that

began c. 1900 and continued into the mid-1930s (Holmlund 1993; Klingbjer and Moberg 2003; Klingbjer *et al.* 2005). Sporadic observations and/or measurements of the glacier's terminus position were made between 1883 and 1981 [e.g. Rabots, 1900; Svenonius, 1910; Karlén 1973; Holmlund and Schytt 1987; unpublished reports, land and aerial photographs (especially F. Enqvist's 1910 photos), and maps] after which more systematic glaciological studies began (Brugger 1992, 2007; Brugger *et al.* 2005; published annual reports and unpublished data from the nearby Tarfala Research Station). Since 1981 an almost continuous mass-balance record exists (available from the World Glacier Monitoring Service, <http://www.wgms.ch>, 5 Aug., 2014). Compiling many of the measurements, Brugger *et al.* (2005) quantified changes in the geometry and volume of Rabots glaciär for the interval 1910–2003. Specifically they

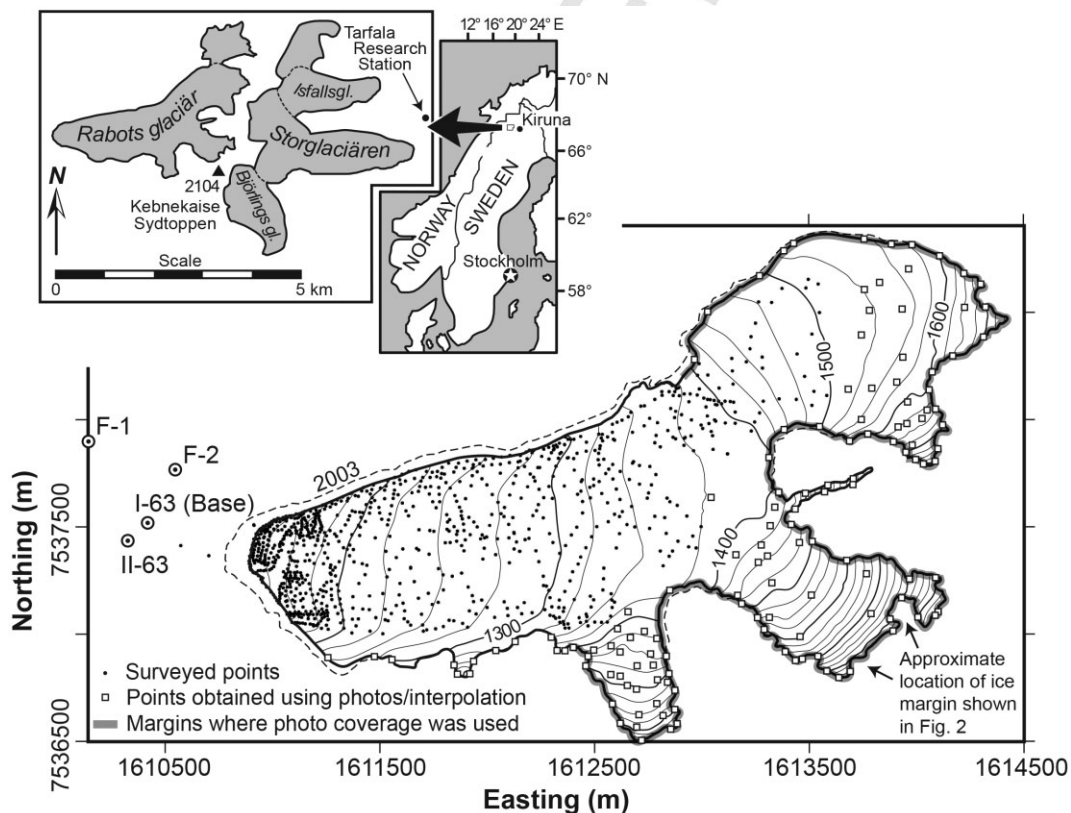


Fig. 1. Location map of Rabots glaciär (inset) and the glacier surface in 2011. Dotted line is the 2003 ice margin. Filled circles show locations of dGPS measurements, some obscured by contours or the line indicating the ice margin. Squares show points obtained by interpolation and photo analyses used to augment the survey data. Coordinate system is RT90 2.5 gon and glacier contour interval is 25 m.

1 show that following an initial sluggish retreat of  
 2  $\sim 2.0 \text{ m a}^{-1}$ , ice recession increased to a relatively  
 3 constant rate, within the resolution of the observa-  
 4 tions, of  $10.1 \text{ m a}^{-1}$  (standard deviation,  $\sigma =$   
 5  $\pm 2.6 \text{ m a}^{-1}$ ) between 1933 and 2003. Although  
 6 changes in glacier volume during this time were  
 7 more difficult to assess owing to temporally sparse  
 8 data, significant ice loss apparently began prior to  
 9 1959. For the period 1959–2003 glacier volume had  
 10 decreased by  $75.9 \times 10^6 \text{ m}^3$ , however the rate of  
 11 volume loss shows a steady decrease from  
 12  $\sim 2.4 \times 10^6 \text{ m}^3 \text{ a}^{-1}$  between 1959 and 1980,  $\sim 1.2 \times$   
 13  $10^6 \text{ m}^3 \text{ a}^{-1}$  between 1980 and 1989, to  $\sim 1.0 \times$   
 14  $10^6 \text{ m}^3 \text{ a}^{-1}$  between 1989 and 2003. Regional  
 15 increases in winter precipitation and its effect on  
 16 glacier mass balance (Pohjola and Rogers 1997;  
 17 Fealy and Sweeney 2005; Kohler *et al.* 2006) might  
 18 provide an explanation for the decreasing rate of  
 19 volume loss. Nevertheless while slowing, volume  
 20 loss continued to be significant up to 2003, leading  
 21 Brugger *et al.* (2005) to conclude that Rabots  
 22 glaciär had not yet completed its response to the  
 23 earlier climatic warming, presumably due to the  
 24 glacier's long response time (discussed subse-  
 25 quently) quantified as either a volume timescale  
 26 (Jóhannesson *et al.* 1989) or as a length response  
 27 time (Oerlemans 2001). Given the potential com-  
 28 peting influences of increased winter precipitation  
 29 and more recent, or accelerated warming in the  
 30 region (Klingbjer and Moberg 2003; Callaghan  
 31 *et al.* 2010) on the mass balance of Rabots glaciär,  
 32 a remaining question is how these are manifest in  
 33 changes in glacier length and volume post-2003.

## 34 **Methods**

### 35 *The 2011 dGPS survey*

36 The 2011 dGPS survey used a real-time kinematic  
 37 technique that was identical to that used in 2003  
 38 (~~carried out 15–21 Jul.~~). A base station receiver  
 39 (Trimble 4600LS) was located at a fixed point (I-63  
 40 on Fig. 1) on the crest of a moraine  $\sim 0.5 \text{ km}$  from  
 41 the ice margin. That fixed point was surveyed in  
 42 2003 and those coordinates were entered as the  
 43 precise location for the base, thus linking the two  
 44 dGPS surveys. After initialization of the base  
 45 station, the ice margin and surface elevations were  
 46 surveyed by a backpack-mounted roving receiver.  
 47 Points were measured individually (as opposed to  
 48 continuously) to ensure that the height of the  
 49 receiver above the glacier's surface was constant.  
 50 Corrections to the rover's position were contin-  
 51 uously broadcast in real time via radio link from the

base receiver. At each location, three measurements  
 were automatically made, averaged, and recorded.  
 Quality control was preset to exclude measurements  
 that exceeded 0.05 m in horizontal and vertical pre-  
 cision, however it should be noted that local surface  
 irregularities (due to differential ablation, meltwater  
 channels, etc.) could in places be several decimeters  
 or more. We varied the density of the more than  
 1000 measurement locations (Fig. 1) according to  
 where more significant changes in surface eleva-  
 tions or the ice margin were expected (e.g. the  
 glacier's terminus) or where the local surface was  
 more undulating and less planar. As in 2003, we  
 were unable to measure points in the tributary  
 cirques because of steep and/or crevassed glacier  
 surfaces and the associated danger inherent in  
 glacier travel. Measurements along the southern  
 margin of the glacier's lower reach were also trou-  
 blesome because the nearly vertical walls effective-  
 ly blocked satellite signals. It was therefore  
 necessary to augment the survey data by observa-  
 tions of glacier change in these areas as described  
 subsequently. During both the 2003 and 2011  
 surveys more than 97% of all dGPS measurements  
 were made directly on the ice surface, summer  
 melting having been well underway and quite exten-  
 sive by the dates of these surveys. (This is reflected  
 in the summer balances for 2003 and 2011 that were  
 $-2.27$  and  $-2.11 \text{ m w.e.}$ , respectively, and substan-  
 tially greater than the period average of  $1.67 \text{ m w.e.}$ )

In addition to those of the glacier surface and  
 margin, several fixed (or control) points were mea-  
 sured to facilitate georeferencing and comparison  
 with the 2003 survey. The mean difference in eleva-  
 tion for these points was  $0.090 \pm 0.099 \text{ m}$ ; all eleva-  
 tions measured in 2011 were lower. It should be  
 noted, however, that for the most stable fixed points  
 surveyed (II-63, a large boulder on a flat moraine  
 crest, and F-1 on bedrock; Fig. 1) the differences  
 were only 0.028 and 0.075 m, respectively. The  
 other points were either marked boulders in the  
 glacier's forefield resting on sometimes (less stable)  
 saturated or frozen fine-grained outwash sediments,  
 or in one case an older fixed point established on  
 loose, steeply dipping bedrock so that some subsi-  
 dence and/or movement may have occurred to these  
 points in the intervening eight years. This is also  
 suggested by changes in horizontal positions that  
 can be as much as 0.14 m as opposed to  $\sim 0.01 \text{ m}$   
 for the more stable points. Therefore given the small  
 differences in the elevation measurements, espe-  
 cially those for II-63 and F-1, and the magnitude of  
 irregularities in the glacier surface, no attempt was



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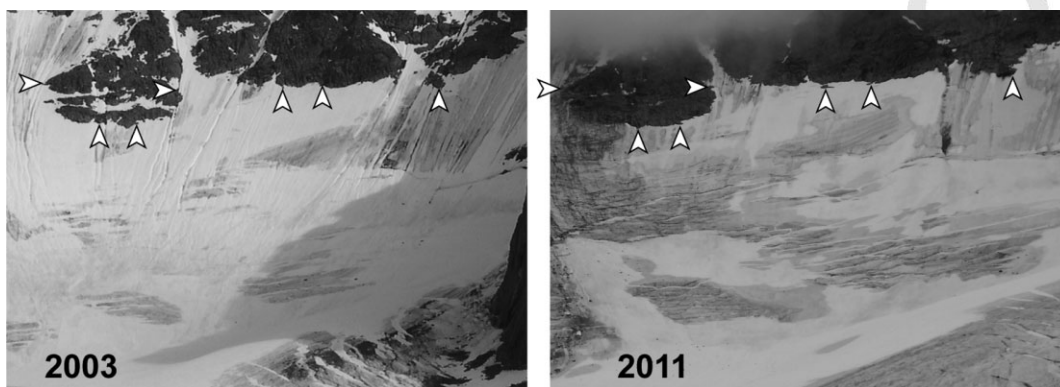


Fig. 2. An example of the photo comparisons used to assess changes in the ice margin in those areas where dGPS surveying could not be carried out due to dangers of glacier travel or lack of satellites. Shown here is part of the southeastern cirque. Arrows show several of the common points used in this comparison. Resolution on originals is far better than that shown here and we estimate that elevation differences greater than  $\sim 1$  m could be detected.

made to correct for the slight apparent discrepancies in fixed point elevations between the 2003 and 2011 surveys.

#### *Augmentation and analysis of survey data*

Supplementary data for those areas where surveying was not possible were obtained by comparing terrestrial photographs from 2003 with those from 2011 (Fig. 2). This was done visually, i.e. we did not perform any ortho-rectification of the photographs. In this way a quantitative assessment was made of the changes in both the position and elevation of the ice margin during this interval. Photographic coverage allowed changes greater than  $\sim 1$  m to be assessed for virtually all of the unmeasured ice margin (Fig. 1) and we note that where revealed by photos very little change occurred. Where the change was more significant, estimates of the coordinates and elevation of the new ice margin were obtained using the 2003 digital elevation model (DEM) and extrapolating the local bedrock slope. Simple linear interpolation of elevation changes was used between the ice margin and the nearest surveyed points with due consideration of persistent trends in surface topography as revealed by maps and photographs. Nowhere was there any indication of thickening. It should be noted that more often than not no discernable changes occurred in the ice margin in these areas while thinning nearby was obvious, but modest. [For comparison, Koblet *et al.* (2010) made a similar observation regarding changes in the accumulation area of nearby Stor-glaciären.] This could suggest a non-linear lowering

of the glacier surface, implying that our interpolation scheme underestimates volume losses in these areas.

The 2011 survey and supplemental data were imported into Surfer (Golden Software) to create a DEM and a contour map of Rabots glaciär. Kriging was used for interpolation to determine glacier surface elevations on nodes over a regular  $10 \times 10$  m grid. Examination of variograms suggested that interpolation using a linear kriging without a nugget effect was appropriate. This explicitly assumes that both measurement error and elevation changes over the relevant distance scale(s) (i.e. nearest neighbors) are small, microtopographic roughness notwithstanding, making the gridding method an exact interpolator.

## **Results**

### *Ice extent and retreat 2003–2011*

The glacier's area in 2011 was  $3.43 \times 10^6$  m<sup>2</sup>, a reduction of  $2.6 \times 10^5$  m<sup>2</sup> (7%) over its 2003 extent (Table 1). The most pronounced changes occurred not only at the terminus but also along the northern margin (Fig. 1). This is not surprising because the southern margin serves as part of the glacier's accumulation area as the steep valley walls enhance snow accumulation through avalanching and possibly wind drift, and reduce solar radiation hence ablation (Brugger 1992). The most significant uncertainties in defining the glacier areas stem from the inability to identify the ice margin where it was obscured by morainal or other debris, both during surveying and in analyzing photographs. This was

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Table 1. Measured changes in the geometry and volume of Rabots glaciär 1910–2011.  $A$  is glacier area,  $\Delta A$  is the change in area,  $\overline{\Delta h}$  is the mean thickness change,  $\Delta V_m$  is the change in glacier volume based on measured changes in surface elevations, and  $dV/dt$  is the rate of volume change.

Year	Retreat rate (m)	Retreat (m a <sup>-1</sup> )	$A^*$ (10 <sup>6</sup> m <sup>2</sup> )	$\Delta A$ (%)	$\overline{\Delta h}$ (m)	$\Delta V_m^\dagger$ (10 <sup>9</sup> m <sup>3</sup> )	$dV/dt$ (10 <sup>6</sup> m <sup>3</sup> a <sup>-1</sup> )
1910	0		4.57				
1933	45	2.0					
1946	197	11.7					
1952	234	6.2					
1959	270	5.1	4.12	-9.8			
1968	358	9.8					
1975	432	10.6					
1980	498	13.2	3.82	-7.3	-12.4 ± 1.5	-51.1 ± 6.0	-2.43
1984	533	8.8					
1989	589	11.2	3.75	-1.9	-2.7 ± 1.9	-10.4 ± 7.0	-1.16
1998	692	11.4					
2003	750	11.6	3.69	-1.4	-3.9 ± 1.9	-14.4 ± 7.0	-1.03
2011	855	13.1	3.43	-7.0	-7.3 ± 0.7	-27.6 ± 2.6	-3.45

\* Referenced to the 1910 maximum LIA extent.

† Data for 1959, 1980, 1989, and 2003 from Brugger *et al.* (2005).

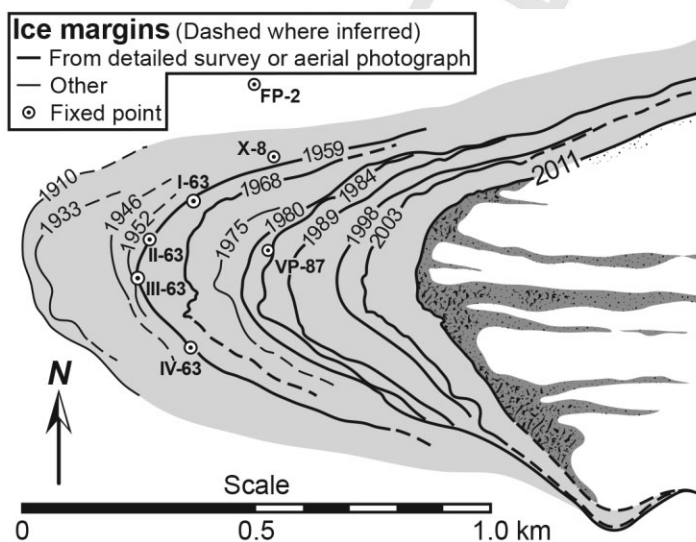


Fig. 3. Ice margins and retreat of Rabots glaciär 1910–2011. Greyed area shows the ice extent in 1910. (After Brugger *et al.* 2005.)

particularly problematic at the southern part of the terminus and mid-reach along the northern margin. In the latter, it was also often difficult to distinguish between (semi-)perennial snow patches and glacier ice. To address these issues, we drew several reasonable alternatives to the ice margins depicted in Fig. 1 and measured glacier area. All areas fell within 1% of one another and we subsequently used this as the uncertainty.

The 2011 terminus geometry is shown in Fig. 3 that documents ice recession from the glacier's LIA

maximum bracketed between 1910 and 1916 (Karlén 1973). [The earlier date is based on the Enqvist photographs but it remains unclear why 1916 was chosen as the older date (P. Holmlund, pers. com., 8 Jan., 2013). For consistency with previous works (Brugger *et al.* 2005; Brugger 2007) we use 1910 when referring to the date of the LIA maximum.] Glacier length between 2003 and 2011 decreased ~105 m based on an average of the difference in 2003 and 2011 terminus positions taken from six lines equally spaced over about 500 m of

1 the glacier's front, this approach being consistent  
2 with measured length changes reported by Brugger  
3 *et al.* (2005). Because of the precision of the dGPS  
4 methodology we assigned an uncertainty of around  
5  $\pm 1$  m for the measured distances that is largely due  
6 to limitations in plotting the survey data and meas-  
7 uring on that plot. The mean rate of retreat for this  
8 interval was  $\sim 13.1 \pm 0.1$  m  $a^{-1}$ .

### 11 *The 2011 glacier surface*

12  
13  
14 Figure 1 shows the contoured glacier surface for  
15 Rabots glaciär in 2011 from the constructed  
16 DEM. To assess the quality of the interpolation,  
17 five random samples of 100 surveyed point eleva-  
18 tions were compared with interpolated values.  
19 The mean residual (difference) in elevations for  
20 these samplings was in all cases less than 0.20 m.  
21 Residuals were normally distributed (standard  
22 deviations  $\leq \sim 2.0$  m), and thus interpolated values  
23 are equally likely to slightly underestimate as  
24 overestimate surface elevations. We thus assumed  
25 an uncertainty of 0.2 m for the glacier surface  
26 where elevations were measured. If, as noted  
27 above, in those areas where measurements could  
28 not be made, surface lowering followed a non-  
29 linear trend (e.g., fitted using a second-order poly-  
30 nomial); thinning in these areas might be  
31 underestimated by  $\sim 1$  m. Considering that these  
32 are approximately one-third of the glacier's total  
33 surface area in 2011, this placed a lower bound on  
34 the uncertainty (via weighting) associated with  
35 the surface of  $\sim 0.5$  m which we then took as the  
36 upper bound as well.

### 39 *Changes in ice thickness and volume 2003–2011*

40  
41  
42 Changes in ice thicknesses  $\Delta h$  were obtained by  
43 subtracting the 2003 DEM of the glacier surface  
44 elevations from those of 2011 (Fig. 4a). Mean thick-  
45 ness change  $\Delta h$  was  $-7.3$  m. Based on the standard  
46 deviation ( $\sigma = \pm 5.5$  m), over most of the glacier the  
47 change in surface elevation was between  $-1.8$  and  
48  $-12.8$  m. As expected the largest changes occurred  
49 near the terminus ( $\sim -27$  m) and over the lower  
50 reach of the glacier. Accordingly, the volume  
51 change  $\Delta V_m$  determined by measured changes in  
52 surface elevations between 2003 and 2011 was  
53  $-27.6 \times 10^6$  m<sup>3</sup> corresponding to a mean rate of  
54  $-3.5 \times 10^6$  m<sup>3</sup>  $a^{-1}$  (Table 1).

55  
56  
57 The uncertainties in these estimates were difficult  
58 to precisely quantify for several reasons, perhaps  
59 the most obvious being the estimation of presumed

60  
61  
62 thinning in the areas where dGPS measurements  
63 could not be made. We therefore evaluated the  
64 uncertainties in the following ways. An uncertainty  
65 of  $\pm 0.5$  m was used for the 2003 ice surface based  
66 on an analysis similar to that described above.  
67 Propagation of uncertainties in the surface eleva-  
68 tions and in glacier area yields an uncertainty  
69 of  $\pm 0.7$  m (or  $\sim \pm 10\%$ ) for  $\Delta h$  and  $\pm 2.6 \times 10^6$  m<sup>3</sup>  
70 for  $\Delta V_m$  (also  $\pm 10\%$ ).

71  
72  
73 When possible, we attempted to explicitly quan-  
74 tify the magnitude of potential errors that might be  
75 introduced as a result of our assumptions and/or  
76 methodologies in order to ensure that these fell well  
77 within the uncertainties stated above. First, even in  
78 the unlikely case that thinning in those areas that  
79 went unmeasured was underestimated by 2 m, an  
80 additional volume loss of  $\sim 2.2 \times 10^6$  m<sup>3</sup> might have  
81 occurred. Second, the kriging algorithm also  
82 resulted in several small areas where minor thick-  
83 ening is indicated (Fig. 4a), and this is most likely  
84 not the case. Eliminating those “offending” values  
85 by arbitrarily setting any positive  $\Delta h$  to zero  
86 changed  $\Delta V_m$  to  $-28.3 \times 10^6$  m<sup>3</sup>, a 3% change over  
87 our best estimate. Alternatively applying a low-pass  
88 filter ( $3 \times 3$  Gaussian) to the gridded  $\Delta h$  values  
89 eliminated all positive values (Fig. 4b), making  
90  $\Delta h = -7.6$  m and  $\Delta V_m = -28.7 \times 10^6$  m<sup>3</sup>. Our uncer-  
91 tainty in  $\Delta V_m$  therefore encompasses those inherent  
92 in each of the foregoing as well as accounting for  
93 irregularities in the glacier surfaces, measurement  
94 inconsistencies (e.g. lean of roving receiver pole),  
95 slight differences in the 2003 and 2011 dGPS eleva-  
96 tions of control points, and/or instrumental error.

### 97 **Discussion**

#### 98 *Comparison of geodetic and glaciological* 99 *mass balances*

100  
101  
102 Systematic measurement of the mass balance of  
103 Rabots glaciär using the glaciological methods  
104 began in 1981 and continues to the present, but  
105 details of the measurement protocols used over this  
106 period have varied. Currently about 100 probings  
107 are made for measuring accumulation (winter  
108 balance) and several (5–10) ablation stakes are  
109 monitored to determine the summer balance (P.  
110 Jansson, pers. com., 8 Aug., 2012). It is therefore  
111 important to evaluate the effectiveness of the  
112 ongoing mass-balance program. We do this by com-  
113 paring geodetic and glaciological (mass balance)  
114 derived volume changes for the interval 2003–11.

115  
116  
117 Unfortunately at the time of this writing, mass-  
118 balance data for the years 2003–04 and 2006–07 are

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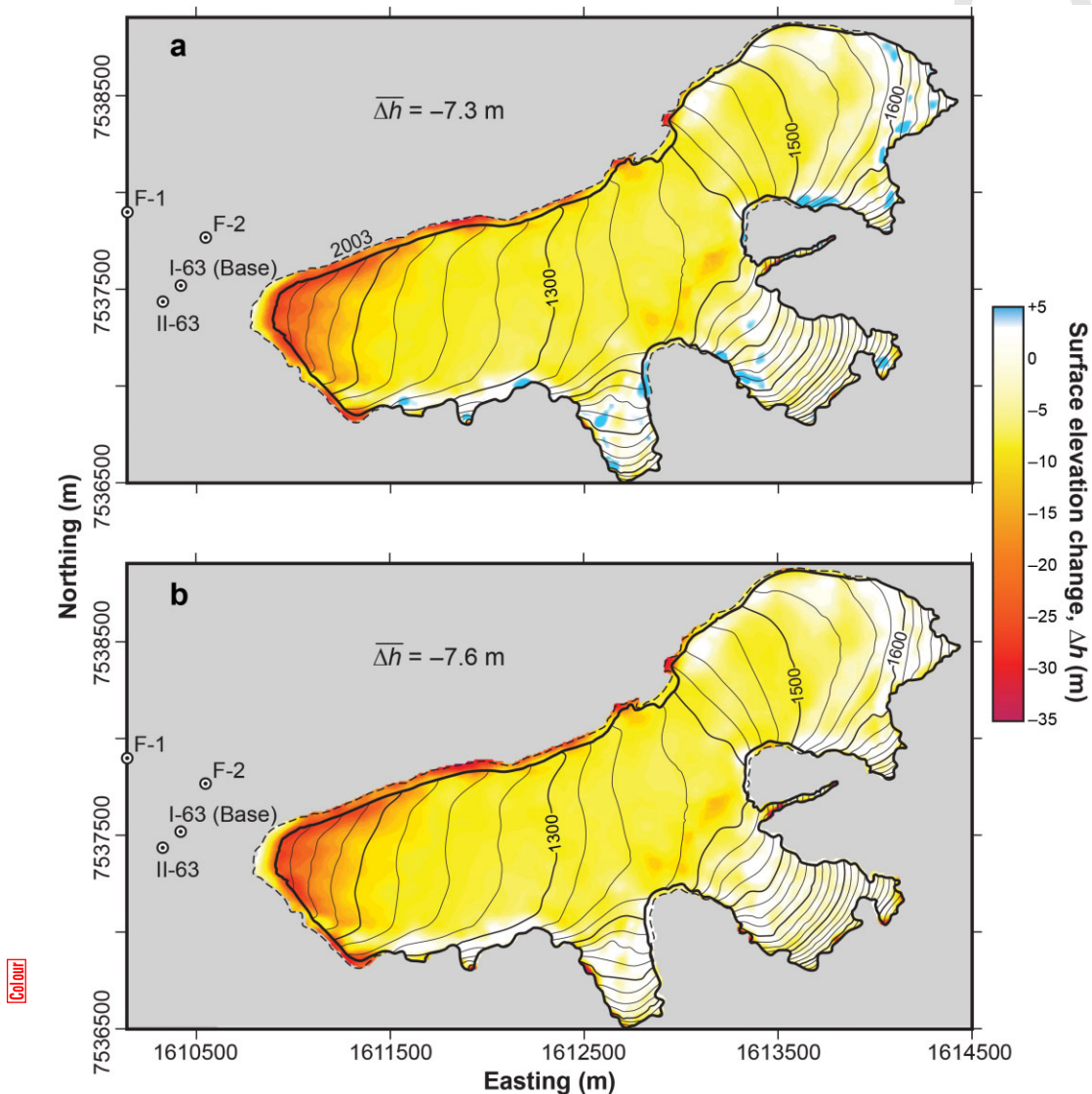


Fig. 4. Image map showing the changes in glacier surface elevations 2003–2011 using (a) unfiltered gridded data and (b) after applying a  $3 \times 3$  Gaussian filter.

not available. However, a good correlation exists between the glacier-wide, annual mass balance  $b_n$  of Rabots glaciär and that of nearby Storglaciären ( $n = 29$ ,  $r^2 = 0.75$ ,  $p < 0.001$ ) that then allows variations in  $b_n$  for Rabots glaciär to be filled in and/or extended, albeit crudely, back to 1946 (Fig. 5a, b). Using this composite record the volume change based on mass-balance variations between 2003 and 2011 is calculated by

$$\Delta V_b = \frac{1}{\bar{\rho}} \sum_{i=2003}^{2011} b_{n_i} A_i \quad (1)$$

where  $\bar{\rho}$  is the mean density of mass (snow and/or ice) lost or gained during the balance year  $i$ ,  $A_i$  the glacier area, and  $b_{n_i}$  the glacier-wide annual mass balance. The density is weighted quantity based on the areal distribution of snow/firn ( $\sim 550 \text{ kg m}^{-3}$ )



and ice ( $916 \text{ kg m}^{-3}$ ) over the glacier's extent using an equilibrium-line altitude that can be related to the annual mass balance by correlation ( $n = 21$ ,  $r^2 = 0.76$ ,  $p < 0.001$ ), and hence accumulation area

via hypsometry. Snow/firn density is based on (1) measurements which show that although typical snow densities range from  $\sim 350$  to  $400 \text{ kg m}^{-3}$  at the approximate end of the accumulation season they increase to  $\sim 400$ – $550 \text{ kg m}^{-3}$  very early in the ablation season (unpublished reports of the Tarfala Research Station); and (2) a mean of  $550 \pm 30 \text{ kg m}^{-3}$  for 50 density measurements (both snow and firn) taken in eight pits during the middle of the ablation seasons in 1983 and 1984 (Brugger 1992). Accordingly this yields a cumulative volume change for 2003–11 of  $-28.1 \pm 2.6 \times 10^6 \text{ m}^3$  where the stated uncertainty includes those in area ( $\pm 1\%$ ),  $b_n$  [ $\pm 0.1 \text{ m w.e.}$ , this being based on analyses by Brugger (1992); cf. Jansson (1999)] and density ( $\pm 50 \text{ kg m}^{-3}$ ). No attempt was made to account for the slight mismatch of the beginning and ending dates of the balance interval and those of the dGPS surveys. The volume loss that might have occurred after the 2003 survey and before the beginning of the 2003–04 balance year included in our estimate of volume change is thought to be comparable to that occurring after the 2011 survey and before the end of the 2010–11 balance year that is not included in our results.

The values of  $\Delta V_m$  and  $\Delta V_b$  are in excellent agreement (Table 2). What is notable here is that a fairly complete mass balance record is available for 2003–11, and this is the only interval wherein  $\Delta V_m$  is determined using dGPS techniques and thus eliminates the uncertainties associated with map analyses [see Brugger *et al.* (2005) for a brief discussion]. The agreement of  $\Delta V_m$  and  $\Delta V_b$  furthermore implies that the present, rather modest program on Rabots glaciär is sufficiently accurate in determining the glacier's mass balance and documenting volumetric changes.

To explore the robustness of the relationship between  $\Delta V_m$  and  $\Delta V_b$  for the earlier three intervals we use values for the former reported previously by

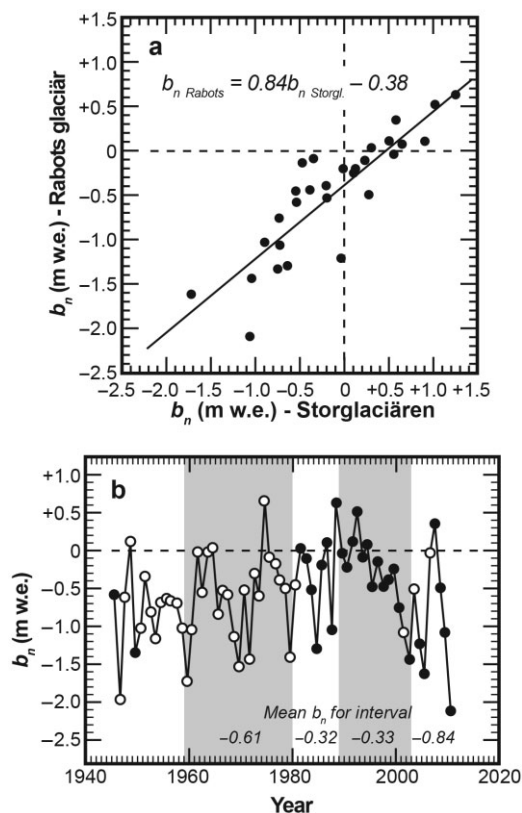


Fig. 5. (a) Correlation of measured specific net mass balances  $b_n$  for Rabots glaciär and Storglaciären ( $n = 29$ ,  $r^2 = 0.75$ ,  $p < 0.001$ ) and (b) the extended record of variations of  $b_n$  for Rabots glaciär for 1946–2011 based on measured values (solid circles) and estimates (open circles) using the regression equation in (a). Mass-balance data are from unpublished updated compilations provided by P. Jansson (personal communication, 2012).

Table 2. Changes in the volume of Rabots glaciär based on measured changes in surface elevations  $\Delta V_m$ , those estimated using a mass balance approach  $\Delta V_b$ , and mean specific net, winter, and summer balances for select intervals between 1959 and 2011.

Interval	$\Delta V_m^*$ ( $10^6 \text{ m}^3$ )	$\Delta V_b^\dagger$ ( $10^6 \text{ m}^3$ )	$\bar{b}_n^\dagger$ (m w.e.)	$\bar{b}_w^\dagger$ (m w.e.)	$\bar{b}_s^\dagger$ (m w.e.)
1959–1980	$-51.1 \pm 6.0$	$-61.0 \pm 6.0$	-0.61	1.00	-1.61
1980–1989	$-10.4 \pm 7.0$	$-14.1 \pm 2.0$	-0.32	1.13	-1.45
1989–2003	$-14.4 \pm 7.0$	$-22.0 \pm 3.0$	-0.33	1.26	-1.59
2003–2011	$-27.6 \pm 2.6$	$-28.1 \pm 3.0$	-0.84	1.09	-1.93

\* Data for 1959, 1980, 1989, and 2003 from Brugger *et al.* (2005).

† Values might use, in part or in whole, a reconstructed record (see text for discussion).

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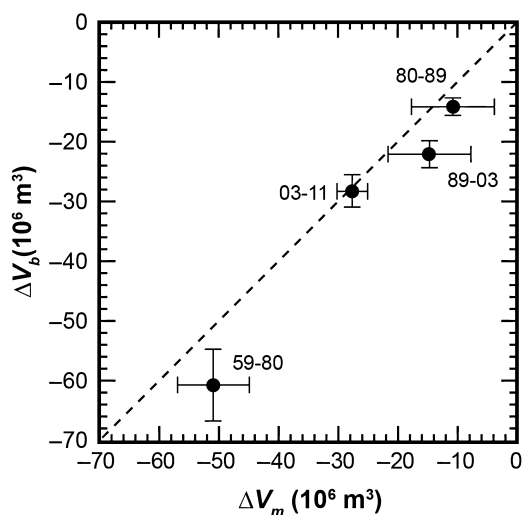


Fig. 6. Comparison of estimates of  $\Delta V$  derived from measured changes in glacier surface elevations ( $\Delta V_m$ ) with those derived using a mass-balance approach ( $\Delta V_b$ ). Error in  $\Delta V_b$  values is taken at  $\pm 10\%$  based on quantifiable uncertainties, but is likely to be larger. Diagonal dashed-line indicates complete concordance. Numbers refer to intervals associated with each data point.

Brugger *et al.* (2005) and determined the latter using Eqn (1). Figure 6a and Table 2 show that in general the agreement is reasonably good, although it might suggest that either the mass-balance approach overestimates, or the values based on elevation changes underestimate the actual volume loss, or some combination of these. However, further comparison of  $\Delta V_m$  and  $\Delta V_b$  is complicated when considering the steps sometimes necessary to estimate  $b_n$ , differences in measuring protocols (i.e. number of stakes, aerial coverage, etc.), potential inaccuracies in the maps, among other things. For example,  $b_n$  values for 1959–80 in particular are entirely based on correlation with those for Storglaciären. It is therefore difficult to say which method more accurately records volume changes for Rabots glaciär prior to 2003.

*Longer-term trends in ice retreat and volume loss*

The current rate of ice retreat ( $13.1 \pm 0.1 \text{ m a}^{-1}$ ) is comparable to the mean rate of  $10.2 \text{ m a}^{-1}$  ( $\sigma = 2.6 \text{ m a}^{-1}$ ) of a trend that began *c.* 1933 (Table 1; Fig. 7a). Terminus behavior is largely controlled by longer-term variations in ice flux from upglacier caused by dynamic adjustments made in response to climatically induced changes in mass balance. The lag in propagation of flux variations to

the terminus gives rise to a glacier's reaction time, or length response time (Oerlemans 2001). However, on shorter timescales local mass and energy balances, among other factors, can affect terminus behavior. For example, a particularly severe melt season(s) could result in substantial ice loss at the terminus that depending on its geometry could appear as an interval of anomalous retreat. Conversely inordinate snow accumulation could mitigate, albeit temporarily, retreat. In an attempt to distinguish between these factors, both the magnitude and rate of ice recession for the 12 intervals during the period 1910–2011 (Table 1; Figs 3 and 7a) were compared with mean summer (June, July, August) temperatures for the same intervals either recorded at, or reconstructed for, the Tarfala Research Station located less than 10 km away from the glacier (Fig. 7b). (It bears mentioning here that temperature measurements in the Rabots valley only exist for the summer months of 1984 and for a few weeks in 2003. These measurements correlate very well with those at Tarfala, and suggest mean temperatures in the Rabots valley are on the order of  $0.5^\circ\text{C}$  lower.) No correlation exists between the magnitude or rate of ice recession and temperature ( $r^2 = 0.04$  and  $0.09$  respectively). Similarly the magnitude or rate of ice recession for these intervals is not correlated to the mean of the corresponding annual mass balance ( $r^2 = 0.07$ ). In addition, neither are these correlated to glacier-wide mean winter balance  $b_w$  ( $r^2$  for both being  $\sim 0.2$ ) that is used here as a surrogate for snow accumulation at the terminus. (Given the typical linear nature of the winter-balance profile over the glacier's lower reach, accumulation at the terminus would approximately scale with  $b_w$ . Note that winter balances for Rabots glaciär prior to the 1981–82 balance year were again obtained via regression of existing values with those from Storglaciären;  $n = 27$ ,  $r^2 = 0.69$ ,  $p < 0.01$ ; no data for two years.) Thus shorter-term terminus behavior does not seem to be unduly influenced by variations in local energy and mass balances. Nor did any changes in bed or ice surface slopes (both relatively constant) affect terminus behavior during retreat from the glacier's maximum extent. (The possible exception here being the glacier's retreat over a preexisting moraine during the 1950s.) This suggests that the terminus is responding predominantly to the glacier's longer-term dynamic response to a mean (summer) warming trend of  $\sim 0.004^\circ\text{C a}^{-1}$  defined by 1910–2010 composite temperature data shown in Fig. 7b, or alternatively  $\sim 0.005^\circ\text{C a}^{-1}$  for only the 1946–2010 temperatures

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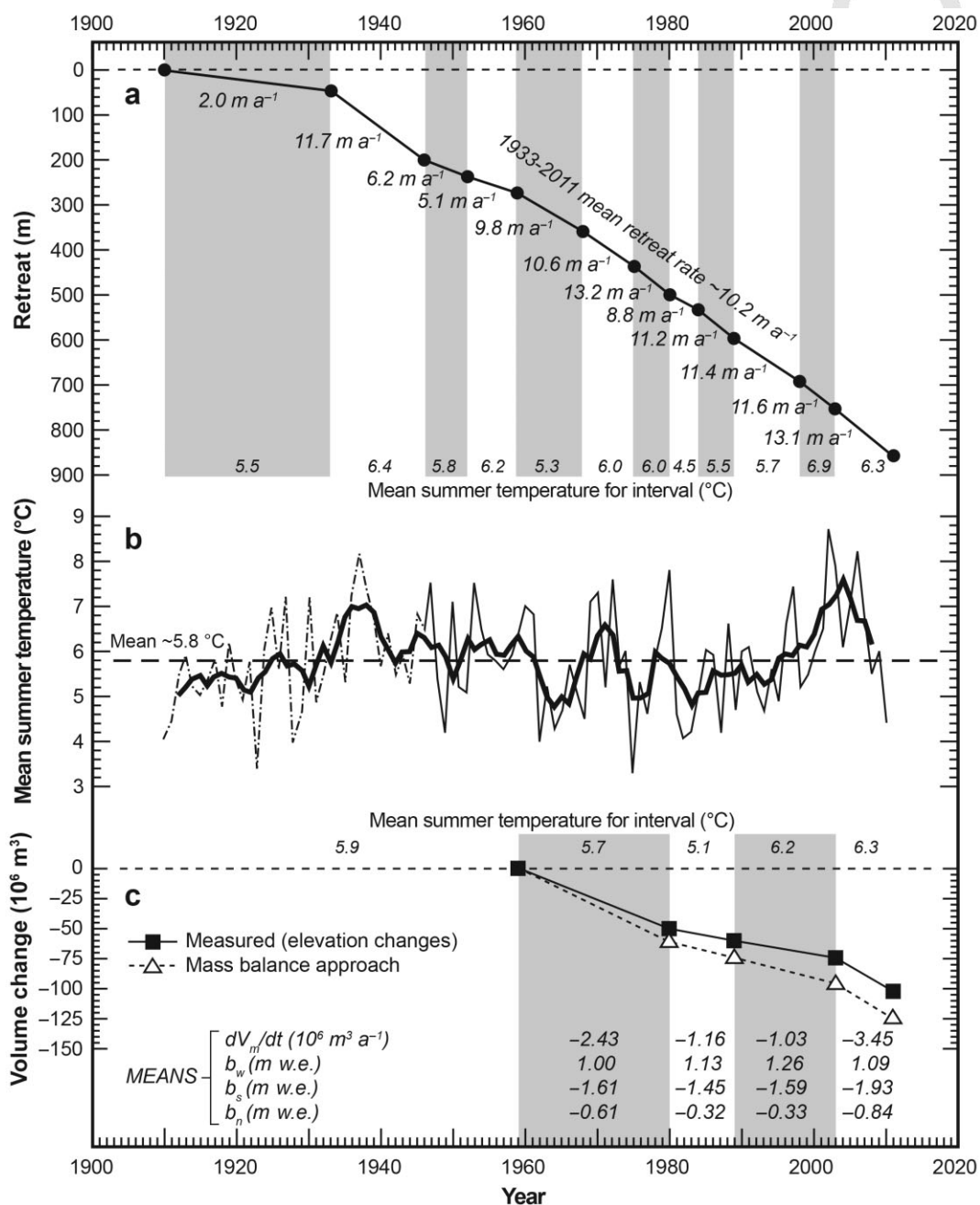


Fig. 7. (a) Terminus retreat and rates of retreat for Rabots glaciär 1910–2011. Gray bars distinguish measurement intervals. (b) Composite temperature record for Tarfala Research Station. Temperature data from 1946 to 2010 (solid thin line) were recorded at the station. Earlier temperature data (dashed line) are based on a correlation between the Tarfala Research Station temperatures and a reconstructed temperature series for Tornedalen ( $n = 57$ ,  $r^2 = 0.73$ ,  $p < 0.001$ ; see Klingbjör and Moberg (2003) for details). Smoothed temperature variation (thick line) uses an averaging window of 5 years. (c) Volume changes based on geodetic measurements, mass-balance data, and a volume-area scaling. Gray bars distinguish intervals. Pertinent data for each interval are also summarized. Error bars are omitted for clarity.

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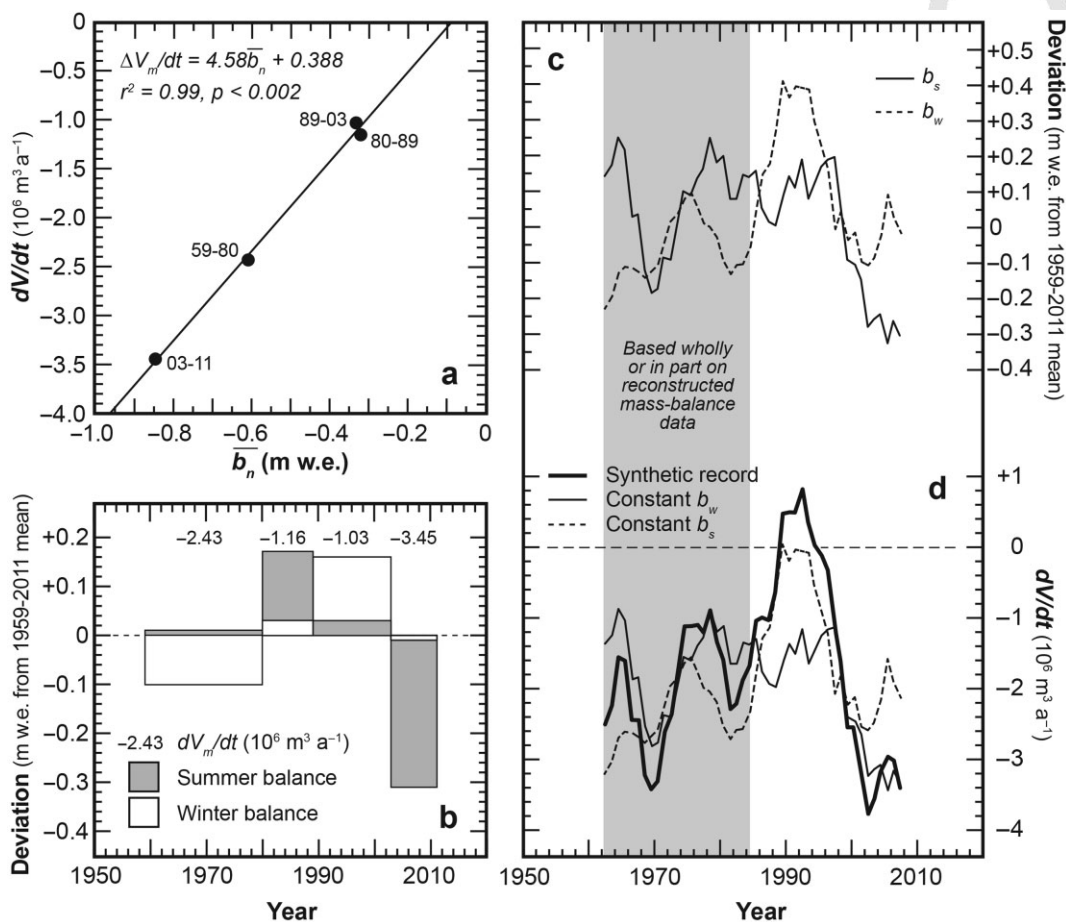


Fig. 8. (a) Correlation of the rate of volume change  $dV/dt$  based on measured changes in surface elevations and mean specific mass-balances for the intervals of 1959–80, 1980–89, 1989–2003, and 2003–11. Numbers adjacent to data points indicate the interval. (b) Deviation of mean specific winter and summer balances for the same intervals from their 1959–2011 mean values. (c) Deviation of yearly specific winter and summer balances from their 1959–2011 mean values and (d) resulting synthetic record of  $dV/dt$  compared with  $dV/dt$  expected if either winter or summer balances were constant (see text for discussion).

recorded at Tarfala. This is not surprising in view of the glacier's long length response and volume response times that are respectively 105 and 215 years (Brugger 2007). As noted previously, the glacier's long-term retreat might be somewhat moderated by the apparent increase in winter precipitation that is reflected locally by a slightly increasing trend of winter balances of Storglaciären ( $-0.006 \text{ m a}^{-1}$ ) and Rabots glaciär ( $-0.005 \text{ m a}^{-1}$  using a composite record of  $b_w$ ), for the 1945–2011 observational interval.

Figure 7c also shows the longer-term trend(s) in volume change for Rabots glaciär. Since 1959, measurements indicate glacier volume was reduced by  $\sim 103.5 \times 10^6 \text{ m}^3$ , or as much as  $125.2 \times 10^6 \text{ m}^3$

based on the mass-balance record. As expected, the rates of change in ice volume for discrete intervals are closely related to variations in glacier mass balance. Rates of volume change  $dV_m/dt$  for 1959–80, 1980–89, 1989–2003 and 2003–11 show a very good correlation ( $n = 4$ ,  $r^2 = 0.99$ ,  $p < 0.005$ ) with the mean annual net balances  $\bar{b}_n$  computed for the corresponding intervals (Fig. 8a; Table 2) and furthermore corroborates the contention regarding the accuracy of the current mass balance program on Rabots glaciär.

More significantly and in contrast to terminus retreat, it is clear that an accelerated rate of volume loss of  $\sim 3.5 \times 10^6 \text{ m}^3 \text{ a}^{-1}$  began after 2003. This rate is appreciably larger than any that can be docu-



1 mented for earlier intervals and is approximately  
 2 double the 1959–2003 average of  $1.7 \times 10^6 \text{ m}^3 \text{ a}^{-1}$ .  
 3 This increase occurred during an interval character-  
 4 ized by a sharp reduction in mean net balance result-  
 5 ing from significantly less-than-average summer  
 6 balances  $b_s$  in combination with winter balances  $b_w$   
 7 (Fig. 8b) that appear to have been quite typical of  
 8 those between 1959 and 2011. Greater-than-  
 9 average winter balances for the previous interval of  
 10 measurement (1989–2003) in combination with  
 11 slightly less negative summer balances resulted in  
 12 a substantially lower rate of volume loss of  
 13  $-1.03 \times 10^6 \text{ m}^3 \text{ a}^{-1}$ . (Interestingly, a notable differ-  
 14 ence between the mean summer balances for 1989–  
 15 2003 and 2003–11 exists despite the mean summer  
 16 temperature being virtually the same, reflecting the  
 17 inability of temperature alone to represent the full  
 18 energy balance involved in ablation processes.) For  
 19 the interval 1980–89, the mean winter balance  
 20 shows a small positive deviation from the 1959–  
 21 2011 average, but the rate of volume loss during this  
 22 period remained low ( $-1.16 \times 10^6 \text{ m}^3 \text{ a}^{-1}$ ) in con-  
 23 junction with considerably less negative summer  
 24 balances. Despite unremarkable differences in  
 25 summer balances with respect to the mean, an  
 26 overall deficit of snow accumulation between 1959  
 27 and 1980 (possibly more characteristic of drier con-  
 28 ditions at that time?) brought about a rapid rate of  
 29 volume change of  $-2.4 \times 10^6 \text{ m}^3 \text{ a}^{-1}$ . This under-  
 30 scores how relatively modest, short-term changes or  
 31 departures in summer balances and/or winter bal-  
 32 ances can significantly impact the rate of volume  
 33 change, all else being equal.

34 To better quantify the relationship and sensitiv-  
 35 ity of variations in  $dV/dt$  to those in winter and  
 36 summer balances we use

$$\frac{dV}{dt} = 5.07\bar{b}_w + 4.41\bar{b}_s - 0.45 \quad (2)$$

37 derived by multiple linear regression ( $n = 4$ ,  
 38  $r^2 = 0.99$ ,  $p < 0.1$ ), where the overbars indicate  
 39 interval means. As before, summer balances prior to  
 40 1981 again obtained via correlation with those on  
 41 Storglaciären ( $n = 27$ ,  $r^2 = 0.64$ ,  $p \ll 0.001$ ). Equa-  
 42 tion (2) allows us to separate the effect of variations  
 43 of winter and summer balances (Fig. 8c) on  $dV/dt$   
 44 by assuming a constant summer and winter balance  
 45 respectively (“controls”). This analysis is also moti-  
 46 vated by the question of whether the latest calcu-  
 47 lated rate of change in volume (2003–11), that  
 48 seems quite high for what can be documented,  
 49 might be unprecedented. Toward these ends, we

calculate values of  $dV/dt$  averaged over an eight-  
 year moving window for 1959–2011 (the period for  
 which *measured* volume change data exist) using  
 the 1959–2011 mean of the control ( $1.10 \text{ m w.e.}$  for  
 $\bar{b}_w$  and  $-1.62$  for  $\bar{b}_s$ ) and the window mean for the  
 variable. We specifically use an eight-year smooth-  
 ing window because this corresponds to the length  
 of the 2003–11 measurement interval, but it bears  
 mentioning that our results are qualitatively the  
 same for other window sizes (e.g. 5- and 10-year).

Despite the shortcomings in using regression-  
 derived values for mass balance components prior  
 to 1981, the results (Fig. 8d) suggest that a sus-  
 tained interval of volume loss comparable to that of  
 2003–11 might have occurred during the late 1960s  
 and early 1970s. This period was characterized by  
 both less positive winter balances and more nega-  
 tive summer balances with respect to their mean  
 1959–2011 values (Fig. 8c), and Fig. 8d suggests  
 that both components contributed to volume loss at  
 this time (i.e. neither alone predicts the expected  
 rate of volume change). An interval of positive  
 values for  $dV/dt$  (volume increase) between the late  
 1980s and the mid-1990s can be largely attributed  
 to substantially increased winter balances aided in  
 part by somewhat less negative summer balances.  
 Part of this interval is reflected in measured volume  
 changes between 1989 and 2003 when the lowest  
 rate of volume loss can be documented. Apparently  
 an overall trend of increasingly negative rates of  
 volume change began prior to the 2003–11 meas-  
 urement interval. It appears this *initially* coincides  
 with both a rather sharp decline in winter accumu-  
 lation – from the significantly higher-than-average  
 amounts in the preceding years – and increasingly  
 negative summer balances. More significantly,  
 however, Fig. 8d suggests that the *continuation* of  
 this trend to the more recent high rate(s) of volume  
 loss, both measured and derived through Eqn (2),  
 can be attributed almost entirely to persistent –  
 arguably extraordinarily so – very negative  
 summer balances. To a large degree, these balances  
 are undoubtedly related to the higher temperature  
 during this time as regression of  $b_s$  on mean  
 summer temperature reveals a significant inverse  
 relationship between these two variables ( $n = 27$ ,  
 $r^2 = 0.54$ ,  $p \ll 0.01$ ) but here again temperature  
 alone fails to fully account for ablation.

Given this longer-term context of glacier behav-  
 ior the recent increase in the rate of volume loss, if  
 continued, might mark the glacier’s response to  
 accelerated warming evident in northern Sweden  
 for the years 2000–10 (Callaghan *et al.* 2010). The

work presented here and previously (Brugger *et al.* 2005; Brugger 2007) clearly demonstrates that Rabots glaciär is nowhere near steady state. In particular, Brugger (2007) derived a response time of ~215 years for the glacier. Even if approximate, this would suggest that Rabots glaciär is sufficiently along in its dynamic response to the earlier warming (*c.* 1900) and thus the rate of volume change might be expected to be decreasing as a new steady state is approached. Instead the magnitude of volume loss has increased, and this appears to be a consequence of the glacier's response to the most recent temperature increase superimposed on its ongoing response to the earlier warming.

### Conclusions

Successive dGPS surveys reveal changes in the geometry of Rabots glaciär that occurred between 2003 and 2011. The terminus of the glacier retreated 105 m, equivalent to a rate of ice recession of ~13 m a<sup>-1</sup>. The present rate of retreat is comparable to the longer-term mean established in 1933, presumably after an initial lag in the glacier's response to concomitant regional warming and slower reaction of the terminus beginning in 1910 (Brugger *et al.* 2005). Between 2003 and 2011 the glacier's surface was lowered by an average of 7.3 ± 5.5 m. The associated volume loss during these eight years was 27.6 ± 3 × 10<sup>6</sup> m<sup>3</sup>. Volume loss estimated by considering the glacier's mass balance for the same period is 28.1 ± 3 × 10<sup>6</sup> m<sup>3</sup>. The agreement of these two estimates provides some confidence that current mass-balance measurements (and program) are adequate in documenting changes in the glacier's mass and volume. Volume loss between 2003 and 2011 occurred at a mean rate of 3.5 × 10<sup>6</sup> m<sup>3</sup> a<sup>-1</sup>. Unlike the glacier's retreat, volume (hence mass) loss appears to have accelerated after 2003, the measured rate of change being substantially greater than any time in the past for which it can be quantified. This pronounced increase in volume loss is clearly linked to more negative summer balances and is thought to reflect the glacier's response to a recent temperature increase (Callaghan *et al.* 2010) superimposed on its ongoing response to the earlier warming. Future research will focus on numerical modeling designed to test this hypothesis.

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