

Surge dynamics in
the Nathorstbreen
glacier system

M. Sund et al.

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Surge dynamics in the Nathorstbreen glacier system, Svalbard

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Abstract

Nathorstbreen glacier system (NGS) recently experienced the largest surge in Svalbard since 1936, and is examined using spatial and temporal observations from DEM differencing, time-series of surface velocities from satellite synthetic aperture radar (SAR) and other sources. The upper basins with maximum accumulation during quiescence correspond to regions of initial lowering. Initial speed-up exceeds quiescent velocities by a factor of several tens of times. This suggests that polythermal glaciers surges are initiated in the temperate area before mass is displaced downglacier. Subsequent downglacier mass displacement coincides with areas of 100–200 times increased velocities (stage 2). After > 5 yr the joint NGS terminus advanced abruptly into tidewater during winter. The advance was followed by upglacier propagation of crevasses, indicating a re-action following from the already displaced mass and extending flow. NGS advanced ca. 15 km, while another ca. 3 km length was lost due to calving. Surface lowering of ca. 50 m was observed in some upglacier areas and during 5 yr the total area increased by 20%. Maximum measured flow rates were at least 25 m d^{-1} , 2500 times quiescence, while average velocities were about 10 m d^{-1} . The surges of Zawadzki breen cycle with ca. 70 yr periods.

1 Introduction

Svalbard is one of the areas with the densest population of surge-type glaciers. Still relatively few glacier surges here have been studied in detail, especially covering both the entire quiescent and surge phase (Murray et al., 2012; Sund and Eiken, 2004). The mechanism behind glacier surging has been studied for several decades, but still the exact cause of glacier surging remains an unsolved question. Remote sensing techniques however, have improved the temporal and spatial data resolution and enables new perceptions on glacier dynamics. In this paper we mainly use satellite remote sensing data acquired between 1992 and 2012 to investigate the surge dynamic of the

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different glaciers in the Nathorstbreen glacier system (NGS), during the largest surge in Svalbard since 1936.

Glacier surges occur as quasi-cyclic rapid velocity accelerations of 10–1000 times the glacier velocity during the longer quiescent phases. Fast flow is associated with a large basal motion component (e.g. Clarke, 1987). The events are triggered by internal rather than external dynamic processes. In the course of the surges large masses of ice are transferred from higher to lower parts of the glacier. During build-up the glacier surface gradient increases. Surges are found both in land- and tidewater terminating glaciers as well as temperate and polythermal ones (Meier and Post, 1969; Dolgoushin and Osipova, 1975; Clarke et al., 1987; Cuffey and Patterson, 2010). Based on studies from the temperate Variegated Glacier, Alaska a hydraulic mechanism, where surges were explained by switches between fast tunnel based basal drainage and slow linked-cavity system, was proposed (Kamb et al., 1985; Kamb, 1987). Some glaciers displaying abrupt speed-up and slow down, while other glaciers go through slower accelerations and decelerations. Differences in observed surge characteristics and dynamics have also lead to a suggestion of at least two separate surge mechanisms between polythermal and temperate glaciers (Murray et al., 2003a). Surges in polythermal glaciers were explained by changes in basal thermal regime which are controlled by a thermal mechanism (Murray et al., 2000; Fowler et al., 2001), while the temperate surges were explained by the hydraulic mechanism (Kamb, 1987). Detailed spatial measurements of surge onset are rare due to the difficulty of prediction. The main emphasis of many investigations has been on the visible part of the surges. For example, several studies have proposed surge initiation during winter (e.g. Raymond, 1987; Echelmeyer et al., 1987; Osipova and Tsvekov, 1991; Eisen et al., 2005; Pritchard et al., 2005). Upglacier propagation of surges from their suggested trigger area has been inferred from upglacier propagation of crevasses (Raymond, 1987; Hodgkins and Dowdeswell, 1994). While mainly focusing on relative changes in surface elevation rather than traditional surge characteristics, Sund et al. (2009) found several early stage

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situations before they had reached the visible and/or advancing stage, and suggested three stages for the active surge phase development.

Due to the remote location and large spatial extent of the NGS, satellite remote sensing methods are well suited to estimate glacier velocities. In particular, synthetic aperture radar (SAR) satellite sensors are able to penetrate clouds, and the active microwave radar instrument allows observations also during the polar night.

The surge of NGS enables study of several glacier tributaries subject to relatively similar climate conditions. This study elaborates on surge initiation as well as propagation in the studied glaciers of NGS and provides new interpretation of surge behaviour. We examine the evolution of temporal and spatial distribution of velocities combined with surface elevation changes between 1936 and 2008. The main focus is on glacier dynamics prior to and during the transition to the visible, heavily crevasses advancing stage. The characteristics of the individual glacier surges are also investigated.

2 Background

The NGS is located in southern Svalbard and covered an area of ca. 430 km² prior to the current surge. The glaciers flow from ca. 700 m a.s.l. in the eastern basins and are gently sloping westwards for 20–30 km where they enter Van Keulenfjorden at sea level. No previous surges have been observed, yet several historical sources points to such events. Liestøl (1973, 1977) inferred a surge with a maximum extent around 1870 extending past the narrowing in the fjord. The Dunér-Nordenskiöld map from 1861–1864 (Hamberg, 1905) indicate that by then the terminus of the NGS as well as Liestølbreen (Fig. 1) was around 12 km further back in the fjord, than during the advance maximum (Harland, 1997). On Hambergs map (1905) from 1898 an advance had occurred and a calving bay at the terminus indicates a beginning retreat. Folded moraines were also present, and Gripp (1927) took these as indications of glacier advance of these basins. Such formations were later viewed as one of the main characteristic of previous surges (Meier and Post, 1969). Even older maps also provide indications of front fluctuations.

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Although they are not precise, several of the maps back to 1634 presented in Conway (1906) has details showing a narrow passage similar to the current narrow part of the fjord, indicating smaller glacier extent than at present. The maps however, do not give any information on which of the glaciers at the head of Van Keulenfjorden were involved in the advances and retreats.

Between 1898 and 2002 the joint terminuses of Liestølbreen and Nathorstbreen retreated ca. 17.5 km. The retreat rate was greatest around 1958–1961 and corresponded well with maximum water depths (Carlsen, 2004). Retreat is common in Svalbard surge-type tidewater glaciers during their quiescent phase when ice fluxes are low. The joint terminuses of Zawadzki breen (previously spelled Zawadskibreen), Polakkbreen and Nathorstbreen (Fig. 1) started to advance into Van Keulenfjorden during winter 2008–2009, while the first indications of mass displacement appeared already in 2003 (Sund et al., 2009).

3 Data and methods

3.1 Satellite data and aerial photos

NGS is located along the commercial flight route between Longyearbyen and Tromsø, enabling acquisition of detailed images from the air (hereafter these are called images) on clear weather days. Several hundred images taken by different photographers from aircrafts and in field during the period 2007–2012, were collected and utilized in this study. In addition older photographs supplemented this collection. Recent aerial photos from the Norwegian Polar Institute (NPI) (0.6 m resolution) were also employed for comparisons.

In order to detect the front position, we used all available optical satellite images from the Moderate Resolution Imaging Spectroradiometer (MODIS) (250 m resolution), as well as radar satellite data from the European Space Agency (ESA) Envisat Advanced

Synthetic Aperture Radar (ASAR) in wide-swath mode. The northern location of the studied NGS and the polar orbit of the Envisat satellite allows for daily coverage.

Our glacier velocity analysis is based on several different satellite synthetic aperture radar (SAR) sensors. From ESA we used ERS-1 data from the 3 day ice-phase mission in January–March 1992, as well as ESAs Envisat ASAR data from 2002–2011. In addition we used Radarsat-2 data from 2010–2012. The SAR sensors operate with a wavelength on the order of 5.6 cm, with a repeat cycle of 24–35 days. Some of the satellite images were also used as comparisons for estimates of glacier area lost due to calving, which was mainly drawn from the images from air.

3.2 Digital elevation models

New and old maps (Hamberg, 1905; NPI main map series of Svalbard, 1 : 100 000 old and new edition, based on aerial photographs from 1936 and 1990) were studied. We used surface Digital Elevation Model (DEM) differencing to derive geometrical elevation changes. The DEM from 1936 was generated from NPI map contour lines based on oblique aerial photographs; the grid size is 100 m. However, some of the undulations appearing are probably exaggerated as a result of the DEM generation method. In the case of Zawadzkibreen and Polakkbreen the 1936 photographs were taken in the up-glacier direction, which makes the elevation accuracy lowest in the upper part, where it is estimated to be ca. 15 m root mean square (rms). The 1990 DEM was photogrammetrically compiled by the NPI, from vertical aerial photographs, and has a grid size of 20 m. The estimated vertical accuracy of the 1990 DEM is ca. 2 m rms in flat areas, decreasing to ca. 6 m rms on steep slopes. The photo/DEM coverage is missing in the upper part of Dobrowolskibreen and Ljosfonn (Fig. 1). Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) from 2003, spatial resolution 15 m, were utilized to generate a DEM following the same concept as in Sund et al. (2009). ASTER-derived elevations are not sufficient to detect changes in topographic details. For the purpose of this study, however, the mass displacement pattern is the important factor. For our ASTER derived DEMs we estimate the vertical accuracy for an individual

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elevation point to be better than 15 m rms for conditions of sufficient optical contrast in the images used, as supported by an extensive test study in eastern Svalbard (Kääb, 2008). The 2008 SPOT 5 HRS (from Satellite Pour l'Observation de la Terre) DEMs have 40 m resolution and vertical accuracy of 10 m with 90 % confidence on surface slopes less than 20 % (Korona et al., 2009), which apply to most of the glacier surfaces treated here. For topography in the fjord, bathymetric contour lines collected in 1999 and 2001 by the Norwegian Hydrographic Service (Carlsen, 2004), were digitized and extrapolated to the shore line to fill the data gaps.

3.3 Velocity measurements

We applied both SAR interferometry (InSAR) (Goldstein et al., 1993; Joughin et al., 1996) as well as the SAR offset-tracking method (Strozzi et al., 2002; Rignot et al., 2011; Paul et al., 2013), in order to estimate glacier velocities. The InSAR method measures displacement in the radar line-of-sight (LOS) direction, with a precision on the order of mm to cm, while offset-tracking provides information about azimuth and range surface displacements, with a precision on the order of dm to m.

Orthophotos compiled from images were used for crevasse tracking for velocity measurements, compilation of front positions and calving estimates. Velocity calculations as well as estimates on calving were made following the method of Eiken and Sund (2012). Accuracy of orthophotos is estimated 50 m (RMS) for well defined details, and to better than 100 m for interpreted features like front positions.

3.4 InSAR processing of ERS-1 SAR data

Based on two ERS-1 SAR scenes (ascending track 37) from the ice-phase mission (3 day revisit time) from 24 January 1992 and 27 January 1992, we produced a 3 day differential interferogram. The radar looks to the right (approximately east) with an angle of approximately 23.5° from the vertical, and the spatial baseline between the scenes was 37 m. A spatial multilooking of 2 looks in range and 8 in azimuth, provided pixels

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with a ground resolution of ca. 40×30 m. The 1990 DEM was used to remove the topographic phase contribution. The interferograms were unwrapped using SNAPHU software (Chen and Zebker, 2001). An area with exposed bedrock was used to calibrate the InSAR phase. The final unwrapped results were then geocoded to an output grid of 20×20 m. Low coherence areas as well as non-glaciated areas were masked out. We stress that the InSAR method for surface displacement has certain limitations. First, the radar measures displacement in the LOS direction only and sensitivity is thus very low in cases where the actual surface displacement vector is near perpendicular to the LOS. Furthermore, earth observing satellites in near-polar orbits fly in a direction close to North–South, and thus sensitivity to (horizontal) surface displacement in this plane is near zero.

In order to compare surface displacement velocities from different methods, we applied a surface-parallel flow approximation and projected the InSAR LOS velocities onto the downslope direction, using a profile along the glacier.

3.5 Offset-tracking processing of SAR data

We used a cross-correlation based method to estimate range and azimuth offsets between pairs of SAR data acquired in the same geometry. The input SAR data from each sensor was coregistered to reference geometry, and the range and azimuth shifts were estimated by searching for the maximum of the two-dimensional correlation function estimated by using rectangular matching window sizes uniformly distributed over the image frame. The quality of the estimates is provided by the signal-to-noise ratio (SNR) of the measurement. We masked out the points with low SNR and removed outliers by applying a median filter.

The output of the offset-tracking algorithm are geocoded velocity fields, measured in the plane spanned by the range and azimuth vectors (Figs. 5a, b and 6). As with the InSAR the surface-parallel flow approximation was used to project the range and azimuth velocities onto the downslope direction, using a profile along the glacier.

4 Results

Following the aim of this study to investigate the surge dynamics and development of the individual glacier, the spatial and temporal observations are evaluated according to the different glacier branches. All coordinates in figures are UTM 33X. Processing of all SAR data were done with the Norut GSAR software (Larsen et al., 2005).

4.1 Pre-advance observations (mainly surge stage 1 and 2)

On Hamberg's (1905) map considerably larger portions of the moraine on the northern side of the fjord is exposed, indicating more advanced melt than on the southern side. Together with the looped moraine pattern this points to that Liestølbreen surged before NGS. Three sets of folded moraines on NGS appear to originate from Zawadzkibreen and Polakkbreen and indicate that several of the glaciers in NGS also participated in the ca. 1870 advance, with Zawadzkibreen as the final glacier to advance. The straight pattern of medial moraines between Zawadzkibreen and Nathorstbreen still points to a relatively simultaneous advance, likewise as during the current surge. In spite of a possibly smaller advance than during the recent surge, the glacier systems occupied a larger part of the fjord width at that time. The NGS retreat between 2002 and 2008 was ca. 200 m, or 33 m a^{-1} except of an initial, slight advance in parts of Dobrowolskibreen.

4.1.1 Dobrowolskibreen

Dobrowolskibreen was the first glacier in the system to surge (Sund et al., 2009), resulting in activation of the stagnant marginal ice at the northern side of the NGS terminus (Sund and Eiken, 2010). The ERS-1-InSAR 3 day interferograms from January 1992 indicate nucleus of increased velocities just upglacier of the confluence of 0.014 m d^{-1} (5 m a^{-1}), while velocities towards the terminus were lower, at 0.004 m d^{-1} (1.5 m a^{-1}) (Fig. 2a). In 2000 velocities were higher, at 0.980 m d^{-1} (358 m a^{-1}) also towards the terminus (Fig. 2b).

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In 2003, when the Dobrowolskibreen was in stage 2 with thinning in its higher elevations and thickening close to the terminus (Sund et al., 2009), large, transversal crevasses were seen in the upper keyhole shaped basin on the ASTER. Comparisons of images from ASTER (between 2005 and 2006) and from aircrafts, shows that ice along the northern margin was activated all the way to the terminus by 2006. Yet, detailed observations were needed to extract the slight advance over several years prior to 2009, appearing in about one third of the entire NGS terminus width. Late in 2007 velocities had increased to 2.7 m d^{-1} (958 m a^{-1}) (Fig. 2c) in the lower reaches of Dobrowolskibreen where ice had been thickening (Fig. 3c) and where the terminus now was advancing and calving.

4.1.2 Zawadzkiibreen

Zawadzkiibreen was extending 8 km further in 1936 than in 1990 (NPI map B11 Van Keulenfjorden). In the 1936 aerial photographs (NPI) the surface had crevasse fields and crevassed margins quite similar to the appearance of the glacier surface in 2008, before the advance. This indicates that Zawadzkiibreen experienced a surge around 1936 which did not lead to terminus advance. Thus the elevation changes (1936–1990) in Fig. 3a cover approximately the entire quiescence and build-up. During quiescence up to 60 m accumulated in the upper basins. The transition zone between build-up and thinning indicates the location of the equilibrium line altitude (ELA) during the period 1936–1990. It spans between 350 and 450 m a.s.l. and is estimated conservatively to 450 m a.s.l. (Fig. 3a). The centerline velocities in January 1992 were highest in the area around the ELA (0.03 m d^{-1}) (11 m a^{-1}), while they were about (0.001 m d^{-1}) ($\sim 0.5 \text{ m a}^{-1}$) in the lowermost part (Figs. 2a and 4). The 1992 velocities represent the situation about 10 yr prior to the initial observed mass transfer and 15 yr prior to initial advance, and were utilized to represent quiescent phase velocities.

Between 1990 and 2003 an initial mass displacement occurred from the uppermost basins. An acceleration (Fig. 4d) caused a maximum of 40 m of drawdown at 1–3 km. Moreover 10–20 m of thickening occurred 3–7 km downglacier, which is below the ELA

(Fig. 4b), and lead to stage 1 representing surge initiation (Fig. 3b). The uppermost areas correspond to the region with greatest accumulation during quiescence (Fig. 3a). In midwinter 2000–2001 the highest velocities appeared in the higher basins (Fig. 2b), with a more than 100 times increase compared to the 1992 velocities in this area.

By September 2008, just before the terminus advance, the surface drawdown included the entire accumulation area, and was greatest in the western part. Ice thicknesses in the lower part had increased by up to 70 m, while the upper portion had lowered about 40 m. The displaced ice aggregated as a bulge at the confluence with Nathorstbreen (Fig. 3c). From 2007 to 2008, Zawadzkibreen developed increasing crevasse fields. No velocities could be extracted from the upper reaches of the glacier, but an area of accelerated velocities of 4.0 m d^{-1} (Fig. 2c) accompanied the area of ice thickening, which propagated downglacier. The velocity increase here was several hundred times the velocities in this area by 1992. 4 km further downglacier velocities were still low, at $\sim 0.5 \text{ m d}^{-1}$ ($\sim 200 \text{ m a}^{-1}$). Orthophoto measurements for March–September 2009 in the latter area show velocities increases to 8 m d^{-1} , 14 times those in 2007.

Figure 4 shows the changes along the longitudinal profile which starts at the SE basin (Fig. 1). Surface elevations in the upper reaches approached the same levels in 2008 as they were in 1936, approximately in the same surge stage (Fig. 3a). Although the glacier has been thinning in the lower part, the surface slope is rather similar below the ELA in 1936 and 1990 (Fig. 4a), while the slope has increased in the upper part. The annual surface elevation change rate (Fig. 4b) shows that during surge stage 1 in the period 1990–2003 the change in mass due to downglacier displacement in the uppermost 3.5 km is higher than the annual snow accumulation during the quiescent period (1936–1990). Accounting for the denser ice compared to accumulated snow, the elevation change rate and mass transfer downglacier of ELA during 1990 to 2003 by far exceeded the annual accumulation rate during the years 1936 to 1990. This implies velocity acceleration in this area, while aggregation of mass just downglacier between km 3.5 and 8.0 indicates that velocities have not yet increased accordingly in this area. Hence, the glacier is thickening in the area ca. 2.5 km downglacier of the ELA

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(Fig. 4b and c). When the surge had reached late stage 2 by 2008, just prior to the advance, the uppermost 8.4 km had lowered, while the lower ca. 8 km has increased up to 70 m. Comparison between Fig. 4b and c shows accelerating mass redistribution from stage 1 (1990–2003) to stage 2 (2003–2008). Velocity changes along the same longitudinal profile show corresponding patterns (Fig. 4d). By 1992 velocities were highest in the upper part, yet no more than 0.03 m d^{-1} (11 m a^{-1}) when measured during winter. As stage 1 was initiated around 2001 levels were at 0.15 m d^{-1} (55 m a^{-1}) in the uppermost reaches, while the area of thickening, between ~ 4 – 11 km , experiences levels up to 0.54 m d^{-1} (200 m a^{-1}). Towards the terminus velocities were somewhat lower, at $> 100 \text{ m a}^{-1}$. Thereafter in 2007 one year before surge advance, the velocities increased with a more mixed pattern, where the highest velocities of more than 4 m d^{-1} are found at 12 – 13 km corresponding to the maximum thickening at this time. In 2010, during advance velocities were $> 7 \text{ m d}^{-1}$, almost twice those in 2007 and had apparently spread upglacier.

4.1.3 Polakkbreen

During the period 1936 to 1990 Polakkbreen thickened by up to 80 m in its higher elevations, with greatest loads on the S-SE part (Fig. 5a). The unchanged area during this period indicates an ELA at 350 m a.s.l. The area with largest accumulation during quiescence corresponds to the zone where a substantial drawdown of up to 60 m is found for the period 1990–2003 (Fig. 5b). A strong surface gradient between thinning and thickening is present in the lower part (Fig. 5a) and corresponds with a slightly curved medial moraine downglacier of the confluence with Nathorstbreen apparent on the 1990 aerial photographs (NPI). This moraine pattern is not present on the 1936 aerial photographs (NPI). During a visit in 1994 (MS) the Polakkbreen-part of the terminus appeared most active of the branches, although still limited. The highest measured velocities on Polakkbreen in January 1992 were 0.043 m d^{-1} (15.7 m a^{-1}) and located along the centerline in the ELA region. Velocities increased steadily upglacier

from the confluence area, while the terminal part had velocities of same magnitude as Zawadzki-breen (Fig. 2a).

On ASTER images from 2003 no crevasses were visible, indicating possible crevasses were limited. Neither on Polakkbreen nor on Nathorstbreen any clearly visible bulge appeared prior to or during advance. Nevertheless an undulating surface developed. Between 2003 and 2008 the uppermost part of Polakkbreen experienced a drawdown of 10–20 m, while the lowermost reaches had the largest thickness increase of ca. 25 m (Fig. 5c). By midwinter 2000 a region in the upper basin had velocities up to 8 m d^{-1} (Fig. 2b) which is an increase almost 200 times the velocities in this area in 1992. Downglacier velocities were 0.38 m d^{-1} (139 m a^{-1}).

In 2008, only limited crevasse fields had emerged on Polakkbreen and Nathorstbreen-Ljosfonn, though more than was visible in aerial photographs from 1961 and 1990 (NPI). Accompanying the advance during stage 3, substantial crevassing appeared in the lower portion of the Nathorstbreen Glacier System (NGS). By end of March 2009, as the terminus advanced, crevasses had spread up-glacier past the glacier confluences. By beginning of September 2009, large crescent crevasses characterized the higher elevations of Polakkbreen.

4.1.4 Other glaciers in the system

In midwinter 2000–2001 velocities in the upper part of Ljosfonn were up to 5 m d^{-1} (Fig. 2b). Detailed studies of images prior to 2008 revealed development of marginal crevasses indicating increased flow, especially along the mountain Keipen (Fig. 1). In late 2007 velocities in Ljosfonn were up to 7.4 m d^{-1} . Downglacier, at the confluence of Polakkbreen, Nathorstbreen and Dobrowolskibreen, they were 2.3 m d^{-1} (Fig. 2c). For the period October 2008 to March 2009 orthophoto measurements showed increases to 8 m d^{-1} at the confluence. In beginning of March 2009, drawdown along the northern sides of the mountain Besshø (Fig. 1) was observed from Ljosfonn, yet the glacier surface still appeared smooth. During summer 2009 crevasses were seen all the way to the upper basins, although they were still limited in the uppermost

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parts. During the following months they increased in amount and size. Orthophoto velocities during winter 2009–2010 (September–February) were 14 m d^{-1} .

Most of the glacier basins draining into Nathorstbreen were involved in the surge. There were however, also basins which were not e.g. Kuvbreen and Blankfjellbreen (Fig. 1). They developed a steep surface slope in the lower part due to surface lowering of the adjacent main trunks. Using the images estimates in some areas (among other Besshø), indicate ca. 50 m lowering of the NGS glacier surface in 2012 as a result of the surge.

4.2 Surge stage 3: advance, frontal velocities and calving of NGS

Envisat ASAR wide-swath images indicated that the abrupt advance occurred after October 2008 where the part of Nathorstbreen draining from Ljosfonn increased its flow simultaneously with Zawadzki breen. The medial moraines in these two branches were squeezed together and appeared as a looped moraine, in front of the slower advancing Polakkbreen. Velocities in this area, calculated from orthophotos over five months (October–March), indicates speeds of $7\text{--}8 \text{ m d}^{-1}$, an increase by a factor of more than three since 2007. A conservative estimate of the front position changes gave winter velocities of 9 m d^{-1} .

As NGS advanced into Van Keulenfjorden during 2009, the glacier front kept its shape and width. This was possibly due to the submarine ridge separating the drainage channel between NGS and Liestølbreen. Between 24 March and 14 May 2009 the front advanced on average 1 km which gives a mean velocity of 20 m d^{-1} . In the period 14 May to 25 September the advance amounted to another 2.5 km, while additional 1 km of the front length was lost due to calving. Velocities were then approximately 25 m d^{-1} .

During winter 2009–2010 from 25 September to 21 February velocities measured from orthophotos were 14 m d^{-1} . The total advance by July 2010 was ca. 10.4 km, excluding calving and including a widening of the terminus into the bay in front of Liestølbreen and Doktorbreen (Fig. 7).

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Around March–April 2010 the front started to splay towards Liestølbreen, widened and started filling the wider head of the fjord. Sea ice was pushed at a distance of ca. 6 km in front of the ice cliff. Between middle and end of July 2010 (Fig. 6a) the frontal 8 km had almost uniform velocities of 12.9 m d^{-1} . Flow rates in the different branches were also rather uniform: Zawadzki breen 5.8 m d^{-1} , Polakkbreen 4.5 m d^{-1} and ice flowing from Ljosfonn and Dobrowolski breen were at levels of 6.1 m d^{-1} . In October 2010 the inner bay was closed off. During the following months (Fig. 6b, c) and winter the velocities lowered in all branches and the lowest velocities in the period were found in March 2011, with maximum velocities of 6.2 m d^{-1} (Fig. 6d). From beginning of June to beginning of July 2011 (Fig. 6e) and in July 2011 (Fig. 6f) velocities increased to 8.8 m d^{-1} in the main trunk. The frontal velocities rose to 9.6 m d^{-1} , about 25 % less than in summer 2010. By August 2011 the NGS had filled the entire inner bay but two smaller areas damming against Liestølbreen and Doktorbreen. The terminus also continued advancing further out Van Keulenfjorden. By end of July to middle of August 2011 (Fig. 6g) velocities slowed to 6.9 m d^{-1} at the front. During October (Fig. 6h) the maximum velocities were reduced to 3.5 m d^{-1} and the main part of the branches were down to 2.4 m d^{-1} . For the period 2010–2012 there are indications of higher velocities in summer than in winter.

In the period August 2011 to August 2012 NGS images show another 2 km advance, indicating average velocities of more than 5 m d^{-1} . Between late March and middle of April 2012 (Fig. 6i) velocities increased slightly compared to October 2011. By autumn 2012 NGS had reached a shallow and narrow part of Van Keulenfjorden and the sea bed in front of the glacier was pushed above the water line. Although some small, new crevasses were observed in situ in the confluence area in August 2012, the overall picture was slow down and crevasse closing processes, which is in accordance with the velocity pattern in Fig. 6. No SAR data were available between May and September 2012. In October 2012 (Fig. 6j) the frontal velocities were reduced to 0.98 m d^{-1} . The main parts of Zawadzki breen and Polakkbreen were at 1.5 m d^{-1} and 1.8 m d^{-1} respectively. Between late October and middle of November (Fig. 6k) as well as during

the next month till middle of December 2012 (Fig. 6I) velocities sustained at the same levels. Compared to the previous winter, flow rates were only about one third those in 2011.

Between August 2012 and August 2013 the advance was another 1.4 km. During the whole period of advance the highest velocities coincide with the region covering the deepest part of inner Van Keulenfjorden (Fig. 7). Altogether the NGS has advanced ca. 15 km, of which some 3 km along at least 4 km width of the terminal part of NGS was lost due to calving during the years 2009–2012, amounting to ca. 12 km². As a result of the surge NGS increased its area by ca. 85 km² or 20 % to totally ca. 515 km².

4.3 Hydrometeorological aspects

In September 2008 anomalously high precipitation and temperatures occurred. NGS is approximately 40 km NE of Hornsund, 100 km SSE of Svalbard Lufthavn, close to Longyearbyen and 200 km SE of Ny-Ålesund (Fig. 1). All these are locations with Norwegian Meteorological Institute weather stations, and the data are shown in Fig. 8. The 30 yr average temperature for Svalbard Lufthavn for September is 0.3 °C and precipitation is 20 mm. Most days in 2008 have higher temperatures than average at all stations. Svalbard Lufthavn displays a more continental climate with slightly higher temperatures, and lower precipitation. Even though the two coastal stations Ny-Ålesund and Hornsund are farthest apart they show the same pattern with anomalously high precipitation for about ten days. The total 2008 September precipitation in Ny-Ålesund was 99 mm, in Hornsund it was 146 mm while Svalbard Lufthavn received 36 mm. Based on location as well as precipitation pattern (Hagen et al., 1993) we assume levels at NGS were closer to those at Hornsund than those measured at Svalbard Lufthavn.

Both in autumn 2009 and 2010 sea ice formed as soon as temperatures dropped below 0 °C in calm weather. We attribute this mainly to freshwater from subglacial drainage and melt from icebergs. On clear weather days MODIS was used to observe possible turbidity of the water in front of the glaciers. In 2009 and 2010 ice mélange prevented observations of possible turbid water close to the terminus. By beginning of

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expressions (Sund et al., 2009). This points to a more frequent cycle of Polakkbreen than ca. 140 yr (1870–2008).

The ELA of Zawadzkiibreen was found to be higher than the 350 m a.s.l. indicated by Hagen et al. (1993), while the ELA of Polakkbreen is consistent with their estimate.

The difference in ELA between the glaciers is consistent with increasing accumulation rates towards east (Sand et al., 2003). Furthermore, cirque shaped basins may provide better conditions for aggregation of snow on lee sides of mountains, causing differential accumulation in the upper part of the glacier. Accumulation on the lee sides was found to be important on glaciers in central Spitsbergen (Jaedicke and Sandvik, 2002; Eckerstorfer and Christiansen, 2011). The regions where the greatest accumulation is seen on Zawadzkiibreen and Polakkbreen (Figs. 3a and 5a) are the same as those that experienced initial lowering (Figs. 3b and 5b). The same pattern was found on the Kroppbreen, Svalbard (Sund et al., 2009). In the lower part of the non-surgings NGS basins the surface slope increased dramatically at the confluence as surface elevation of the main glaciers lowered. This however, did not release a surge, which we attribute to lack of sufficient build-up in these basins. The quiescent phase velocities found for Zawadzkiibreen and Polakkbreen (Fig. 2a) are consistent with other Svalbard surge-type glaciers during quiescence (Nuttall et al., 1997; Melvold and Hagen, 1998; Sund and Eiken, 2004). Comparisons also supports that the slow velocities in the lower parts are evidence of predominantly frozen to bed conditions.

5.2 Mass transfer and velocities during surge (stages 1–3)

On Zawadzkiibreen, the lowering onset in the upper part was accompanied by increased velocities. The fact that mass in the accumulation area is removed faster than it is gained, implies that velocities have risen above balance velocities in this area. Hence ice flux is larger than balance flux, as is defined for surge state. Comparison between Figs. 4a and 4b shows that by 2003 an initial downglacier thickening had already occurred also in the main trunk, although this was not interpreted from a single period (Sund et al., 2009).

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We suggest the following course for the surge development. The areas gaining largest amount of mass during quiescence will reach enhanced gravitational forcing prior to the surrounding areas. (1) In stage 1 this results in locally increased velocity in areas with temperate bed conditions and may be succeeded by several processes:

5 increased frictional heat at the bed, arising from accelerated velocities and generating more subglacial melt water. The amounts of water available at the bed impact on the basal motion (e.g. Iken, 1981; Bartholomaus et al., 2008).

Some crevasses may also open at this or a later stage, allowing more water to penetrate to the bed and thus accentuating the process. (2) An initial drawdown occurs and areas immediately downglacier gain more inflow of ice and hence temporary increased ice thicknesses within a relatively short time. Local short term increase in ice thickness in a new area (stage 2) also influenced by higher levels of subglacial meltwater from further upglacier, leads to an accentuation of the process with increased velocities, increased frictional heat and melt water production as well as possible disruption in the existing basal drainage system. The increased ice flow in additional areas may result in opening of new crevasses, further enhancing the water flow into the glacier. (3) The above outlined process continues and expands over an increasing area. A positive feedback effect is induced and the initial surge nucleus propagates downglacier (Budd, 1975; McMeeking and Johnson, 1986). The effect of the frictional heat will be most important when sliding and basal drag are intermediate (Raymond, 2000). Since the process lasts more than a year it is probable that surface meltwater penetrating into the glacier also contributes to this process. On Variegated Glacier the velocities increased five times those during quiescence, one year prior to the surge advance (Raymond and Harrison, 1988).

A blockage of the subglacial drainage system (Budd, 1975; Kamb, 1987; Eisen et al., 2005), caused by compression between the increasingly activated and the stagnant ice masses, enables the surge nucleus to grow. This occurs through the closure of the basal drainage system during autumn, followed by trapping, and eventually release of water. Englacial water is released, as described by Lingle and Fatland (2003), towards

the end of this stage or early in stage 3. Fast flow could also arise from deformation of the bed which can also accommodate an inefficient drainage system (Harrison and Post, 2003 and references therein). We propose that also in polythermal glaciers the surges are initiated in the temperate area and downglacier propagation is an effect of a chain reaction.

5.3 Glacier surface expressions

A surface bulge was visible prior to stage 3 on Zawadzki-breen. It formed at a stage when the activated upglacier surge nucleus reached what we interpreted to be a thermal dam, where major parts of the ice were frozen to bed, and aggregated there. At this stage the situation at Zawadzki-breen has its equivalent with the situation at the time of the study of Bakaninbreen, Svalbard (Murray et al., 1998). No clearly visible bulge appeared on Polakkbreen or Nathorstbreen. We relate the lack of bulge in some branches to difference in individual glacier properties such as thermal regime. We suggest that the rate of downglacier thickening in polythermal glaciers is governed by the temperature conditions of the downglacier ice as well as differences in sliding conditions. Glaciers with substantial cold margins or large areas of cold basal ice may to a larger extent experience a damming effect. Thus larger downglacier increases in ice thicknesses occurs during the process, such as seen on Zawadzki-breen. While glaciers only moderately affected by a cold bed, have a faster propagation of increased velocities (although still moderate) and less surface elevation changes as their terminuses slowly start to advance. At this stage the advance may be counterbalanced by increased calving. The larger Ljosfonn – Nathorstbreen branch was presumably more affected by the latter. The radio echo-soundings along the centre line of Nathorstbreen by Dowdeswell et al. (1984) (Figs. 1 and 9), indicated ice thicknesses up to 400 m and an up to 200 m thick cold surface layer in the lowermost 18 km, from the terminus position before the surge. On the other hand the causative feedback effect will be lower when ice thickness changes are smaller.

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In previous studies upglacier propagation of velocities and crevasses has been related to upglacier propagation of the surge (e.g. Meier and Post, 1969; Hodgkins and Dowdeswell, 1994; Murray et al., 2003b; Pritchard et al., 2005). We suggest a different interpretation here. Our results show that in stage 3, when upglacier propagation of crevasses occurred, the mass displacement and surge had progressed downglacier for several years. Areas with ice thickness increases during stage 2, subsequently experience surface lowering as the surge propagates further downglacier during stage 3. Hence the upglacier propagation of velocities is not representing an upglacier propagation of the surge from the initiation area, but a re-action in the last stage of the ice displacement. At this stage glacier velocity has reached the highest levels during surge development, leading to a rapid extension of the upglacier ice. The characteristics only reflect a time-section of the entire surge. Yet, the pattern of upglacier propagation will also depend on the individual geometric and thermal properties of the glaciers.

Meier and Post (1969) reported on how surges in tributary glaciers could trigger surges in the main trunk glacier. We argue that this only occurs if the glacier is in late quiescent of early surge stage and is thus “ready to surge”. The surges in the main basins of NGS illustrates the concept of factors reaching a threshold value in a certain order, while basins that do not meet these criteria will not surge, as seen on Blankfjellbreen and Kuvbreen. Both glaciers are located in the same geological formations as the surging branches (Hjelle, 1993). Even though the lower parts were affected by the surging ice, it did not lead to surges in these tributaries. This may be due to frozen conditions at the bed or lack of sufficient build up. The results in this study also show that glaciers in a system may influence each other by muting the magnitude of the surge due to damming effects, but can also enhance the force of the surge if their late quiescence and early surge stage coincide.

When the Dobrowolskibreen surge propagated to the terminus, it possibly “released” the joint terminus of NGS from its frozen northern lateral margin. This did not occur when Skobreen, Svalbard surged into the trunk glacier Paulabreen, the latter was not influenced upglacier (Sund, 2006). Just prior to the Skobreen surge, no mass displace-

ment took place in (upper) Paulabreen, while it had already gone through a partial surge ca. 30 yr before. Another example is Ingerbreen's surge into the confluence Richardsbreen, Svalbard. It affected the joint front, but did not release any surge in the upper part of the latter (Sund et al., 2009). The surge will propagate as long as the driving factors exceed the factors holding the glacier back. Medvezhiy Glacier, Pamirs was reported to have surged before the calculated critical mass was achieved when the damming ice disappeared (Osipova and Tsvekov, 1991).

5.4 Winter initiation of surges?

The advance of NGS started during winter 2008–2009 as an abrupt event. The advance pattern is therefore unlike some other observations of surges in Svalbard (Murray, 2003a), while it has similarities with others (Liestøl, 1969). We emphasize the geometrical development of NGS leading to this event. Yet, the weather event in September 2008 possibly enhanced the velocity acceleration and hence the timing and force of the advance. At the fast flowing Kronebreen, close to Ny-Ålesund (Fig. 1) the weather conditions in September 2008 accounted for 15% of seasonal surface melt and also resulted in greater mean velocities than during summer (Sund, 2011). In Hornsund (Fig. 1) as well, the weather event caused increased glacier velocities (M. Błaszczyk, personal communication, University of Silesia, Sosnowiec, Poland, 2012). Given the common temporal and spatial availability of data for the area a decade or two ago, the abrupt start of the advance of the NGS could easily have been interpreted as an abrupt initiation of the surge. Raymond and Harrison (1985) suggested that the timing of surge initiation in winter coincides with a collapse of the subglacial conduit system due to residual summer meltwater trapped within the glacier. On Monacobreen, Svalbard more than doubled velocities were observed during winter 1991/92 (Murray et al., 2003a). Lingle and Fatland (2003) put forward that certain meteorological conditions are necessary to provide large amounts of water for storage in temperate glaciers and are thus a strong contributor to timing of surge onset. By “surge” they refer to behaviour corresponding to stage 3 (Sund et al., 2009). On the one hand this is in consistence

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with observations of the sudden advance of NGS starting in October. On the other hand, we have shown that the advance of NGS surge was just the final stage in a row of events. The ongoing mass displacement of NGS had by that time lead to some crevassing, enabling water to penetrate through the crevasses towards the glacier bed. Even though the different branches were not exactly at the same surge stage, the September event likely promoted expediting the surge development to stage 3 for some branches. Large mass transfers as well as velocity increases through several years prior to advance, have been demonstrated in this study. Yet, the many reported advances during winter still points to the probable importance of changes of the subglacial drainage system during winter during the preceding stages.

5.5 Front velocities, advance rate and surge termination

The low retreat rate of NGS during the last 6 yr prior to the surge advance occurring at the same time as thinning at the terminus, indicates shallow waters in this region. This is supported by the radio echo profile (Dowdeswell et al., 1984) which started about the same location as the 2008 terminus (Fig. 1). Just prior to the advance of NGS, the front of the Nathorstbreen section had few visible sign of surge behavior. Yet, increased mass displacement (Sund et al., 2009) as well as velocity increases (Fig. 2c) had occurred upglacier for several years, proving that retreated front positions are not evidence of contemporary quiescence. Moreover, this dynamics is consistent with the pattern of downglacier propagation from the upper part of the glaciers, such as on Variegated glacier (Kamb et al., 1985) and differs from the suggested fast flow initiation in the lower part of some other tidewater glaciers (Murray et al., 2003a, b; Pritchard et al., 2005). Nevertheless, if only the last period before advance is considered, the observations of NGS coincide with the latter observations as well.

Velocities between March and May 2009 and May to September 2009 based on front displacement are consistent with those measured by Eiken and Sund (2012) for the period March–September. Indicating velocities reached maximum during summer 2009 when the terminus advanced across the deepest part of the bay. They consti-

tute 100 times front velocities of ca. 0.2 d^{-1} (73 m a^{-1}) measured in the same area in summer 1950, during quiescence (Liestøl, 1977).

The flow rates of NGS during surge are consistent with those observed from other large surges in Svalbard such as Hinlopenbreen (Liestøl, 1973) and Bråsvellbreen (Schytt, 1969) although they did not reach entirely to the levels founds at Negribreen (Liestøl, 1969). They are also comparable with velocities in other regions, for example West Fork Glacier, Alaska (Harrison et al., 1994) and Sortebræ, Greenland (Pritchard et al., 2005).

Both in September 2011 and 2012 and summer 2013 large amounts of turbid water was seen in the fjord in front of NGS. The existence of turbid water has in previous studies been related to a collapse of a linked-cavity drainage system (Kamb, 1987) and hence an abrupt end of the surge (Kamb et al., 1985). By 2012 the surge had been active for at least 10 yr, before flow rates decreased substantially and NGS reached shallow sea bed.

6 Conclusions

Using a combination of DEM differencing and several optical and radar remote sensing methods we have studied the spatial and temporal development of glacier dynamics during surges of NGS. A previous surge of Zawadzki breen was found around 1936 lead to a cycle period of about 70 yr, which also corresponds with the period of previously proposed advance of the NGS by 1870.

Areas of maximum accumulation coincided with the surge stage 1 areas of initial lowering, indicating that the initial trigger area was located in the temperate accumulation area of the polythermal glaciers. These regions were also the first to display increased velocities, accelerating to a factor of more than hundred compared to those found during quiescence. Subsequently in stage 2 mass propagated downglacier with an additional velocity increase 13 times those in stage 1, thus supporting this initial dynamics is a part of the active surge process.

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Possibly the suddenness of the surge advance (stage 3) initiated in October 2008 was encouraged by a weather event in September 2008. This also led to a surge synchronisation of the different glaciers which were in various surge stages. Nevertheless, the surge advance was the final stage in a row of events of increased velocities and mass displacement occurring during several years. During the first months of advance (stage 3) the velocities in the terminal part were ten times those found during stage 2.

Uplacier propagation of crevasses occurred mainly in stage 3 as a reaction to increases in velocities after the initial downglacier mass transfer. The different glaciers displayed variations in their surge surface expressions, which we relate to differences in their thermal regime and geometry. Hence, we hypothesize that surges in polythermal glaciers, landbased as well as tidewater terminating, are initiated well up in the temperate area. The propagation and development of the surge take different forms depending on the extent of cold marginal and frontal ice.

In order to reveal the entire mechanism behind the surge phenomenon it might be necessary to reconsider the study focus. Attention must be increased to include and acknowledge the initial dynamics stages as a part of the surge process prior to the visible surge, in order to catch limited but important changes, although this will probably lead to some redefinitions.

During a 5 yr period NGS lost ca. 3 km length and 12 km² of ice due to calving and advanced ca. 15 km filling the entire head of the fjord. Surface lowering at the order of ca. 50 m was observed in some areas upglacier and the total NGS area was increased by 20%. Maximum measured velocities of ca. 25 m d⁻¹ were found during summer 2009 constituting more than 2500 times observed velocities during late quiescence. Turbid water was observed in the fjord in the last three autumns.

Supplementary material related to this article is available online at <http://www.the-cryosphere-discuss.net/7/4937/2013/tcd-7-4937-2013-supplement.zip>.

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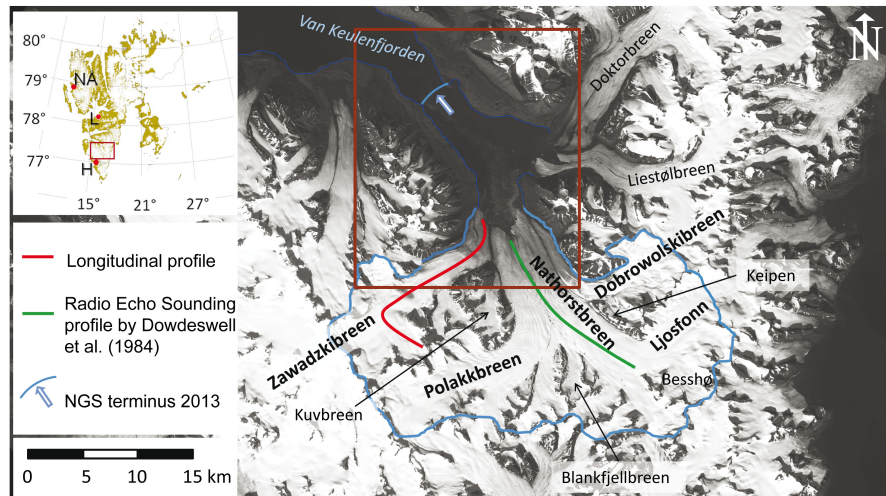


Fig. 1. Inset map: Main islands of Svalbard, with meteorological stations Ny-Ålesund (NÅ), Svalbard Lufthavn – Longyearbyen (L) and Hornsund (H). Nathorstbreen Glacier System (blue outline) is draining into Van Keulenfjorden. Red frame refers to Fig. 7. Background image: 1 September 2008 is from the Système Probatoire pour l’Observation de la Terre (SPOT) Stereoscopic survey of Polar Ice: Reference Images and Topographies (SPIRIT) programme, © Centre National d’Etudes, France (2007) and SPOT image 2008 all rights reserved.

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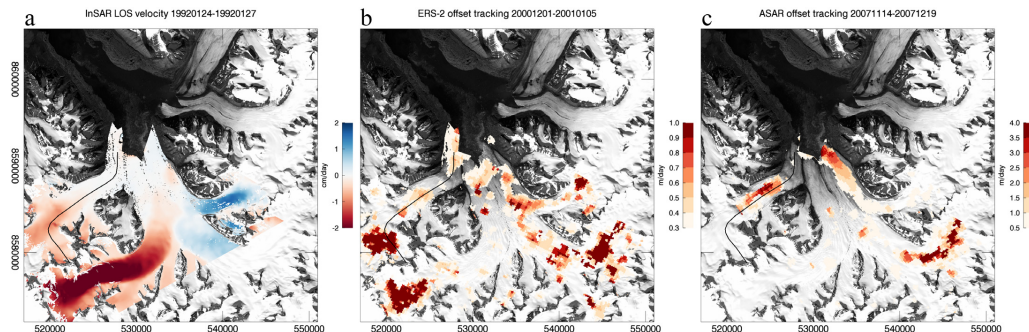


Fig. 2. Temporal and spatial velocity distribution in the Nathorstbreen Glacier System. Profile along glacier apply to Fig. 4. **(a)** ERS-1 3 day InSAR LOS velocity from January 1992. **(b)** ERS-2 35 day offset-tracking from 2000–2001. **(c)** Envisat ASAR 35 day offset-tracking from 2007. Note how increased velocities propagate from higher to lower part of the glaciers. Colour scales are enhanced to display velocity variations. Velocities higher than indicated on colour bars are found in limited areas.

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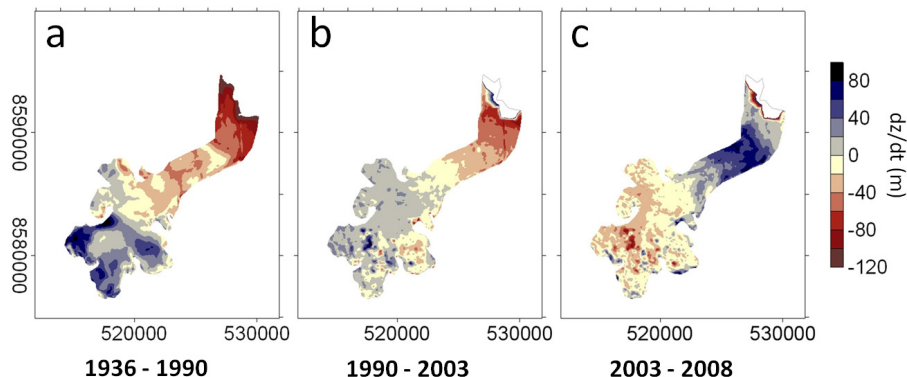


Fig. 3. Surface elevation changes on Zawadzki-breen flowing from SW to NE. **(a)** Build-up in the upper part, thinning and retreat in the lower part. **(b)** Surge stage 1: initial lowering in SE basins. **(c)** Surge stage 2 at the transition to stage 3: lowering of the entire upper part and aggregated thickening in the lower part. Note that the lowermost terminus has not yet advanced and still display thinning and retreat, and also that some of the areas with the heaviest accumulation corresponds to the areas with initial lowering.

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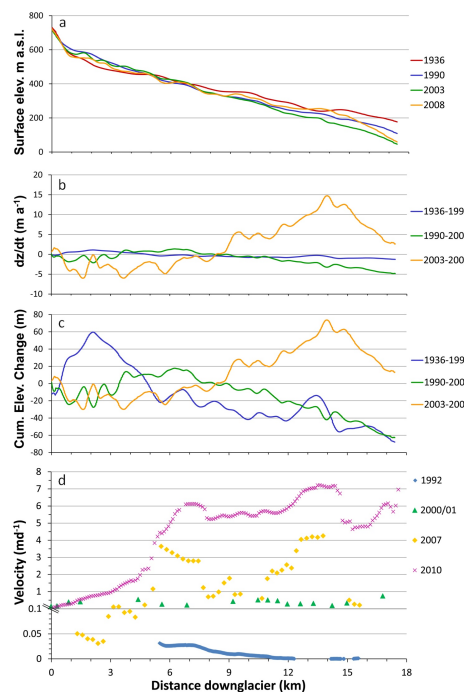


Fig. 4. Zawadzki breen surface elevation changes along the longitudinal profile from SE to N (location in Fig. 1). **(a)** Surface elevations during one surge cycle 1936 to 2008. **(b)** Annual surface elevation change rate: Blue colour indicates quiescent build-up, green colour surge stage 1 and orange colour surge stage 2 (transition to stage 3). Stage 1–3 refers to surge development as described by Sund et al. 2009. **(c)** Cumulative surface elevation change from previous surge in 1936 towards the transition to stage 3 in 2008. Colours correspond to the surface profile at the end of each interval. **(d)** Surface velocities from ERS-1 InSAR (1992), ERS-2 (2000–01), Envisat ASAR (2007) and Radarsat-2 (2010) offset-tracking velocities projected onto the profile assuming surface-parallel flow. Note the break in the vertical scale at 0.1 m d^{-1} .

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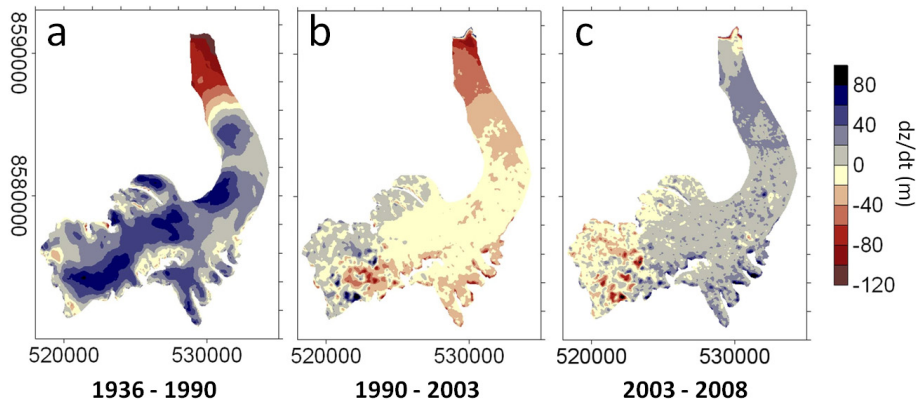


Fig. 5. Surface elevation changes on Polakkbreen flowing from SW to N. **(a)** Build-up in the upper part, thinning and retreat in the lower part. **(b)** Surge stage 2: lowering in the southern part of the upper basin. **(c)** Surge stage 2 at the transition to stage 3: more extensive lowering of the upper part, mass transfer and thickening in the lower part. Note that the lowermost terminus has not yet advanced and still display thinning and retreat, and furthermore that some of the areas with the heaviest accumulation corresponds to the areas with initial lowering.

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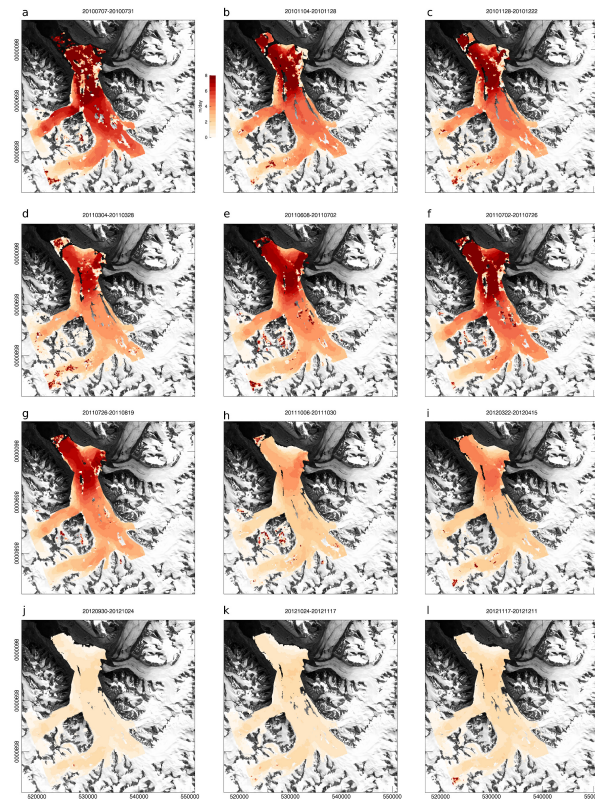


Fig. 6. Surge advance velocity distribution in the different branches of NGS in 2010–2012. Velocities are estimated using offset-tracking of 24 day pairs of Radarsat-2 Fine mode SAR data. The border between the glacier terminus and sea ice appear as a black curved area. Background image by SPOT SPIRIT.

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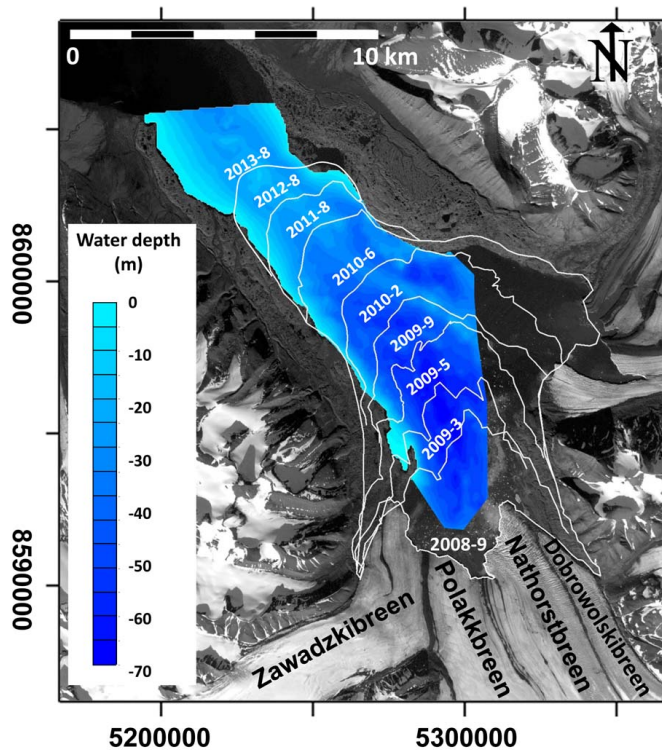



Fig. 7. Front positions (year and month) of Nathorstbreen Glacier System during the advancing stage (3) of the surge. Advance by 2013 is 15 km. Bed topography modified from Carlsen (2004). The minimum depth is 11 m, while the maximum depth is 76 m. Background image from 1 September 2008 by SPOT SPIRIT.

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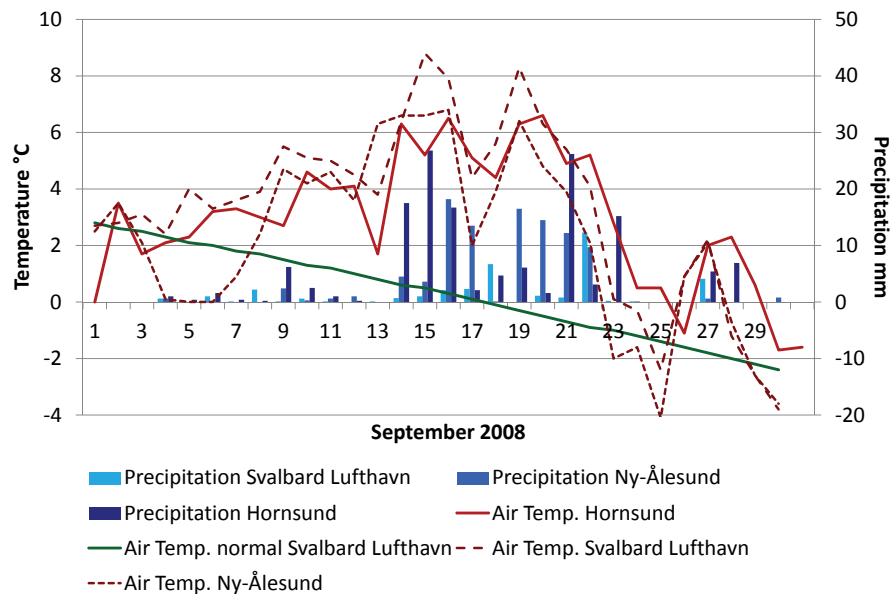


Fig. 8. Daily temperature and precipitation in Ny-Ålesund, Longyearbyen (Svalbard Lufthavn) and Hornsund (location in Fig. 1) in September 2008, just prior to surge advance. Data are from Norwegian Meteorological Institute.

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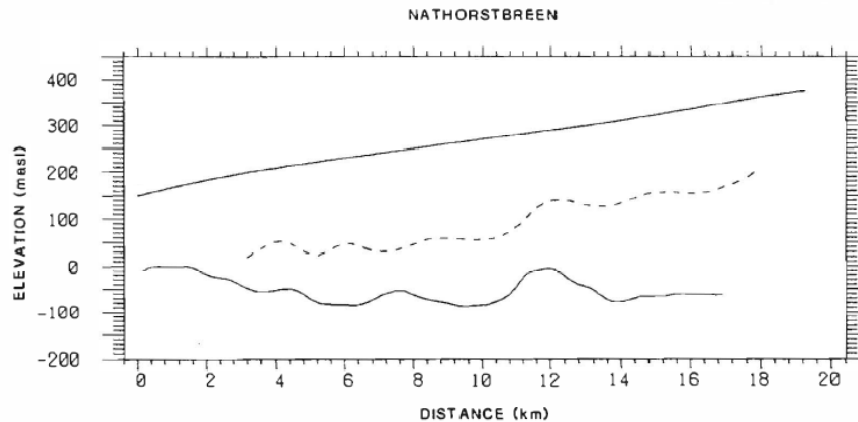


Fig. 9. Radio Echo sounding profile along Nathorstbreen (location in Fig. 1) from Dowdeswell et al. (1984), by permission from the Norwegian Polar Institute.

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