Sustained high winter glacier velocities from brief warm events

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Key Points:

- Thanks to an innovative ice cave monitoring technique we measure water fluctuations in an englacial channel
- A single week-long warm event in the winter can lead to a more than doubling of the velocity of the glacier for more than 3 months

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Abstract

Glacier winter velocities are generally slower than summer velocities, that increase early in the season when meltwater runoff reaches the bed. Velocities generally decrease later in the season as the subglacial system becomes channelized. With an innovative ice cave monitoring technique, here we show that a single week-long warm event in the winter triggers an internal drainage system flooding event, leading to velocity doubling for the remainder of the winter, unlike summer speedups when additional meltwater forms efficient drainage channels that reduce glacier velocity. As the climate warms and surface melt and rain events increase during winter months, sustained high winter glacier velocities are likely to occur more often. Increasing glacier velocity near the terminus leads to additional ice entering the fjord, and an increase of sea level rise contribution during these sustained season-long events.

Plain Language Summary

Most studies of changes in glacier speed focus on summer. However, because of the long Arctic winter season, changes in the winter can cause large changes in annual average speed. The causes of these changes are generally well-understood and linked to water inputs to the inside and the base of the glacier. Unfortunately, even if these changes are well-known they are rarely observed, especially in winter, due to the difficulties of Arctic winter fieldwork, in-situ sub-glacial fieldwork, and installing and recovering sensors at the glacier bed. Thanks to an innovative ice cave monitoring technique, we present six months of winter water fluctuation measurements inside an englacial channel. We show that a single week-long warm event in the winter can lead to a more than doubling of the winter velocity for more than 3 months. In a warming climate, more winter melt and rain is likely to occur, and may lead to increased winter glacier velocity, additional ice entering fjords, and increased sea level rise.

1 Introduction

Glacier velocities are known to be strongly related to their internal drainage system (IDS) hydrology (R. S. Anderson et al., 2004; Bingham et al., 2006; Brinkerhoff et al., 2016; Hooke et al., 1997; Iken & Bindschadler, 1986; Jansson, 1995; Mair et al., 2003; Willis, 1995). This relationship is driven by rain and surface meltwater delivery to the glacier bed (Doyle et al., 2015; Vieli et al., 2004). Thus, the IDS allows the weather and climate information to be transferred from glaciers' surface to their interior (Mavlyudov, 2006). This enables glaciers to react quickly via their dynamics to changes in atmospheric conditions.

The cause of both the seasonal average velocity and short-term velocity variability of glaciers are well understood. During the summer melt season, high temperatures are associated with large water (meltwater and precipitation) input to the IDS. Initially, few small channels exist, the basal system becomes flooded, basal friction in is reduced, and glacier velocity increases. Later, large channels form, basal water is drawn into them and evacuated from the surrounding bed, basal friction increases, and glacier velocity decreases (B. Anderson et al., 2014; Bindschadler et al., 1977; Clarke, 1991; Rippin et al., 2005). From this behavior, more runoff may actually reduce the mean summer velocity (Pimentel & Flowers, 2011; Sundal et al., 2011; van de Wal et al., 2008). Short-term velocity spikes are generally associated with lake drainage, rain, or excessive warm events - all of which can generate sufficient surface meltwater to temporarily overwhelm even mature late-season efficient subglacial channels (Schoof, 2010; B. Anderson et al., 2014; Bartholomaus et al., 2008; Hart et al., 2019). During the winter, due to low temperatures almost no water enters the IDS. Channels creep closed, and increased ice/bed coupling reduces glacier velocity (Cowton et al., 2013; Harper et al., 2005).

Because the system is reset each winter, it is easier to overflow the subglacial system during the winter or early spring, should sufficient water supply exist. Evidence for this comes from surge events, many starting in winter, and correlated with high IDS water pressures (Harrison & Post, 2003; Kamb et al., 1985; Lingle & Fatland, 2003; Sund et al., 2014).

Because most of the crucial processes seem to occur during the melt season, and perhaps because of the difficulties of Arctic winter fieldwork and sensor operation, glacial hydrological and dynamics studies mainly focus on the summer period, although some winter studies do exist (e.g., Hart et al. (2019); Schoof et al. (2014); Sole et al. (2013)). However, in the Arctic the accumulation season is much longer than the ablation season, so the average ice velocity (and therefore discharge), and especially the median ice velocity and discharge, are determined primarily by the winter season. Changes in the winter velocity may therefore have a larger impact than changes in the summer velocity (Harrison & Post, 2003; Kamb et al., 1985; Lingle & Fatland, 2003; Sund et al., 2014).

To address this gap, here we present a time series of 2016/2017 winter observations at a polythemal high Arctic glacier. Data include water level from inside an englacial channel, velocity measurements of the glacier surface, and automatic weather station (AWS) data.

2 Study area

Hansbreen is a 15.6 km long polythermal tidewater glacier, with a mean ice thickness of 171 m (Grabiec et al., 2012), situated at 77° 04'N, 15° 38'E in southwest Spitsbergen (Figure 1). It flows towards the south, extending from sea level to 664 m altitude, with a velocity increasing towards its terminus, ca. 150 m yr^{-1} at the front and of 55–70 m yr^{-1} 3.7 km upstream, reaching a maximum during late spring–early summer (Błaszczyk et al., 2009). It is climatically, enviromentally, and glaciologically similar to other Svalbard tidewater glaciers (Grabiec et al., 2012; Hagen et al., 1993, 2003). In the past, water level measurements have been collected directly from within its moulins (Schroeder, 1998a, 2007; Vieli et al., 2004) and indirectly via ground-penetrating radar (GPR) (Turu, 2012). Hansbreen summer speedup events have been related to increased basal water pressure due to rainfall or summer warm events (Vieli et al., 2002).

Our study focused on measurements from inside an englacial system called Crystal Cave (CC, Figure 1). It is a well-known system that has been active since at least 1967 (Benn et al., 2009; Turu, 2012; Schroeder, 1998a, 1998b). Currently, CC recharge is known to be from four moulins and may be occasionally supplied during high discharge events by one additional moulin and one supraglacial stream (Figure 2) flowing at the interface of the nunatak slope Tuva and the tributary glacier Tuvbreen (Figure 1). A subglacial model (Decaux et al., 2019) shows the presence of a subglacial channel nearby CC's entrance and GPR measurements confirm its connection with the subglacial drainage network (Pälli et al., 2003).

3 Method

3.1 Velocity

Hansbreen velocity is measured with a Global Navigation Satellite Systems (GNSS) receiver that sampled location every 3 hours at stake 4MONIT (Figure 1). We calculate the daily speed based on each midnight positions, and define the "baseline velocity" as the mean velocity from Dec 1 2016 through Feb 1 2017.

The velocity has also been surveyed for decades at a stakes network along Hansbreen's center line (stakes 2 to 11 in Figure 1) but with a much lower temporal resolution. GNSS positions were manually recorded weekly for stakes 2 through 5, and monthly

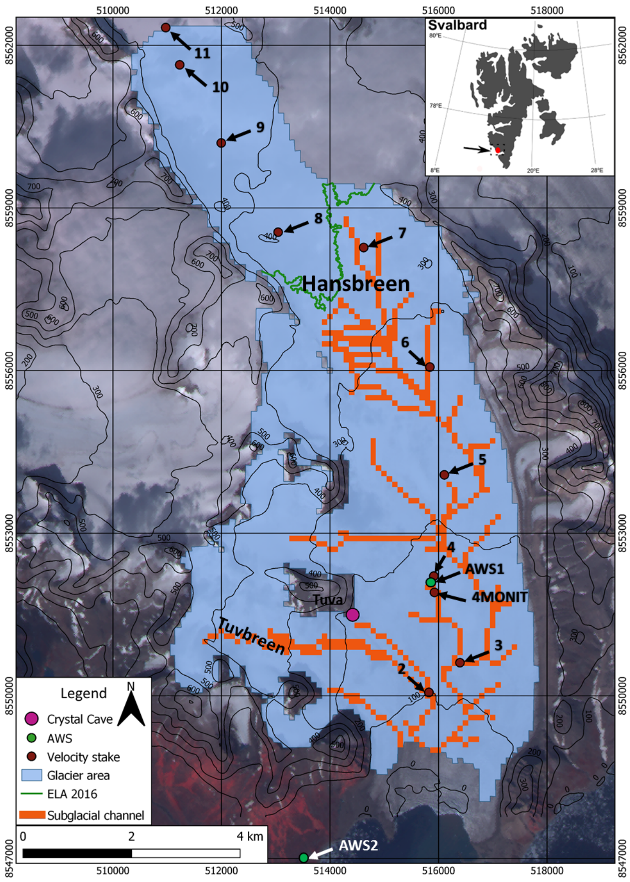

Figure 1. Map showing the locations of the glacier Hansbreen in Svalbard (insert map), the two automatic weather stations (AWS), the 10 velocity stakes, Crystal Cave, subglacial channels from Decaux et al. (2019) and the ELA for 2016 (the accumulation area being above the ELA and ablation area being below the ELA). The background map is a SPOT satellite image acquired on 16 August 1988 and the coordinate system used is WGS 1984 UTM zone 33° N.

for stakes 6 through 11, dependent on weather conditions. The minimum observation times at those stakes is between 20 and 30 minutes. Total error, taking into account GNSS receiver, stake tilt, and human factor is estimated at 20 cm.

3.2 Englacial water pressure

Englacial water pressure was recorded by placing HOBO 250-Foot Depth Water Level Data Loggers in the center of one cave system (Figure 2). Here we report data from CC system near the nunatak Tuva (Figure 1). Data loggers were set to record values every 30 minutes, resampled to daily in post-processing, and have a resolution of 2.55 kPa for a typical error of 3.8 cm water level.

Sensors were placed in the cave by drilling anchor points into the ice above a vertical shaft, then hanging cables down in the center of conduit (Figure 2). Stabilization cables were used to keep sensors from attaching to and freezing into ice walls by manually rappelling down to the sensor and attaching it to three horizontal cables, anchored into the ice walls at ~120 degrees apart. Although we installed two sensors in CC using this method, the upper one was quickly buried due to the cave surface melting down and drifting snow (Figure 2).

Here, we report data from the lower senor installed in CC at ∼100 m total distance from the cave entrance, in ice ∼74 m thick. The sensor was installed 28 m above the glacier (measured) bed and 46 m below the ice surface (estimated) (Figure 2).

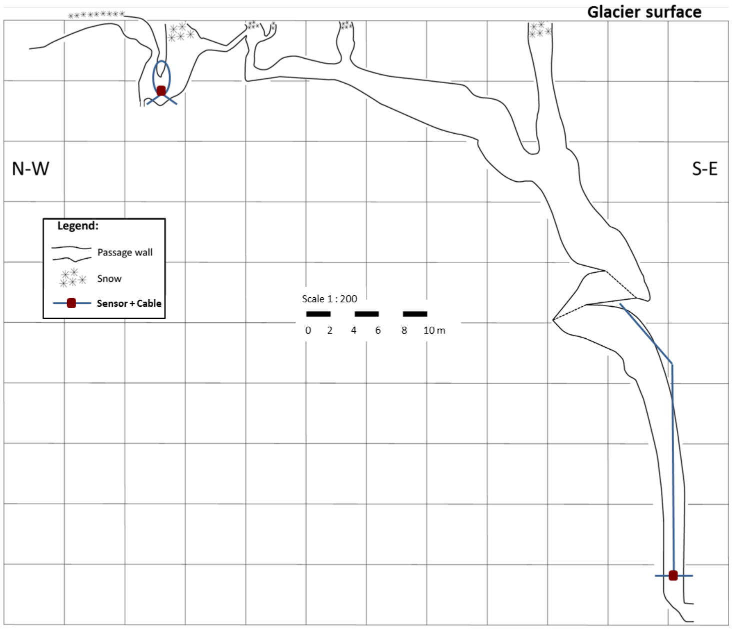

Figure 2. Vertical profile of englacial channel Crystal Cave from April 2017 with sensors locations.

3.3 Weather

A nearby weather station, 1.8 km away from CC, provides air temperature (AWS 1) (Figure 1). Air temperature comes from a Campbell Scientific 107 sensor at ±0.1° C resolution and sampled every 10 minutes, averaged to daily resolution in post-processing. Precipitation measurements were made at AWS 2 (Figure 1) with a multi-type gauge that measured both solid and liquid. Results were into liquid water equivalent in millimeters. Precipitation measurements are slightly offset temporally, with a day defined as beginning at 6 a.m. on the observed day and ending 6 a.m. on the day after.

4 Results

During our study period, from 2016-12-01 to 2017-05-29, daily average temperatures recorded at AWS1 were below 0° C except for two periods with only one longer than two days. From AWS2, both of these warm events included rainfall events.

We highlight the week-long warm period (actually 6 days 20 hours and 30 minutes) with a maximum of 3° C and mean of 1.5° C in early February 2017 (Figure 3).

Prior to this warm event, water level (scale is m above glacier bed) is below the sensor, and therefore not shown on the graph. We can only report that there was < 28 m of water in the englacial system. Beginning 2 days 16 hours and 30 minutes after the first melt day (temperature > 0° C) water rose over 3 days to more than 60 meters above the glacier bed, remained there for more than 9 days, then rapidly dropped below the sensor for about 7 days, and returned to a level around 38 meters above the bedrock (k around 0.55), and remained there for the duration of the record, until June 2017 (Figure 3).

The velocity record begins at $\sim$0.07 m day^{-1} and climbed slowly to just under 0.1 m day^{-1} from December 2016 through February 10 2017. Coincident with the increasing water level, velocity rose above 400 % of the baseline velocity, then reduced to more than 200 % of the baseline velocity and remained there for the duration of the record, until June 2017. The entire ablation area of the glacier reacts similarly to velocity data recorded at 4MONIT stake, but the accumulation area does not (Figure 4).

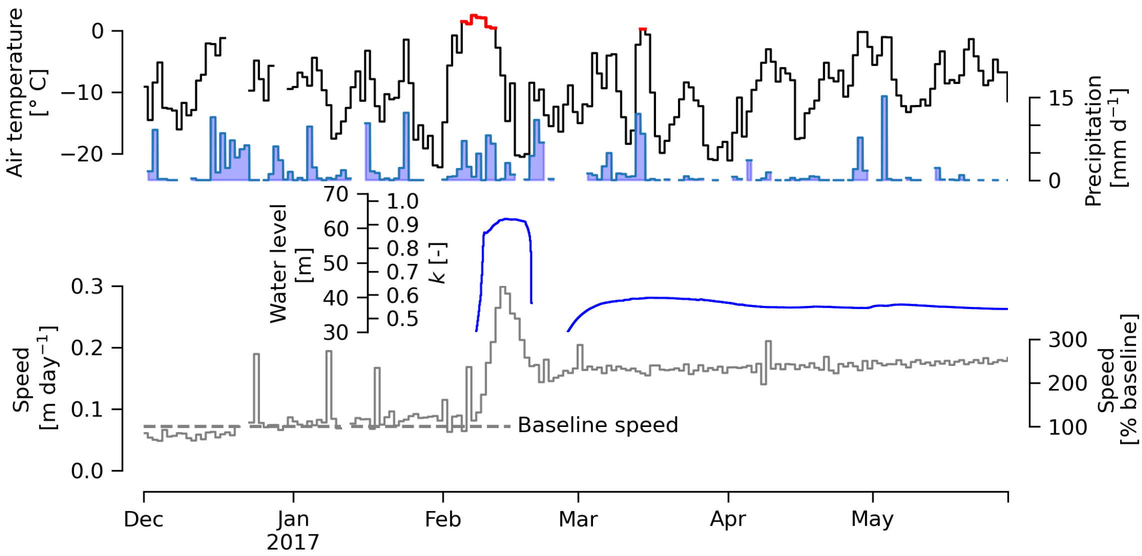

Figure 3. Time series of air temperature at AWS1 (positive red line and negative black line), precipitation at AWS 2 (blue area), water level above the bedrock at CC and flotation fraction (k) (blue line) and glacier speed at 4MONIT stake (grey line) during the winter 2016 / 2017.

5 Discussion

Persistent high winter velocities due to a brief winter warm event represent a dynamic response and potential significant source of sea level rise. Winter warm events may have a more significant influence on the annual average velocity of the glacier than a warm event or a rain event in summer. The occurrence of winter warm events (Graham et al., 2017; Pitcher et al., 2020; Vikhamar-Schuler et al., 2016; Wu, 2017) and winter rain events (Łupikasza et al., 2019; Nowak & Hodson, 2013; Peeters et al., 2019; Sobota et al., 2020) in the Arctic is increasing. Until enough winter water enters the glacier system to cause efficient drainage channels, it is likely that small volumes of winter water will act analogous to a ”spring event”, inducing an increase in ice velocity (Bingham et al., 2006; Kessler & Anderson, 2004; Mair et al., 2003), with no subsequent decrease, as shown here. This increase only occurs in the ablation area (Figure 4), as the IDS does not exist in the accumulation area (Decaux et al., 2019).

After the warm event and coincident rise in water level (Figure 3), the water briefly drops below the sensor (< 28 m above the bed), then returns to 38 m above the bed. From this point onward in our record, the water level generally decreases, while velocity slightly increases. We hypothesize this is due to creep-closure of subglacial channels (Duval, 1977;

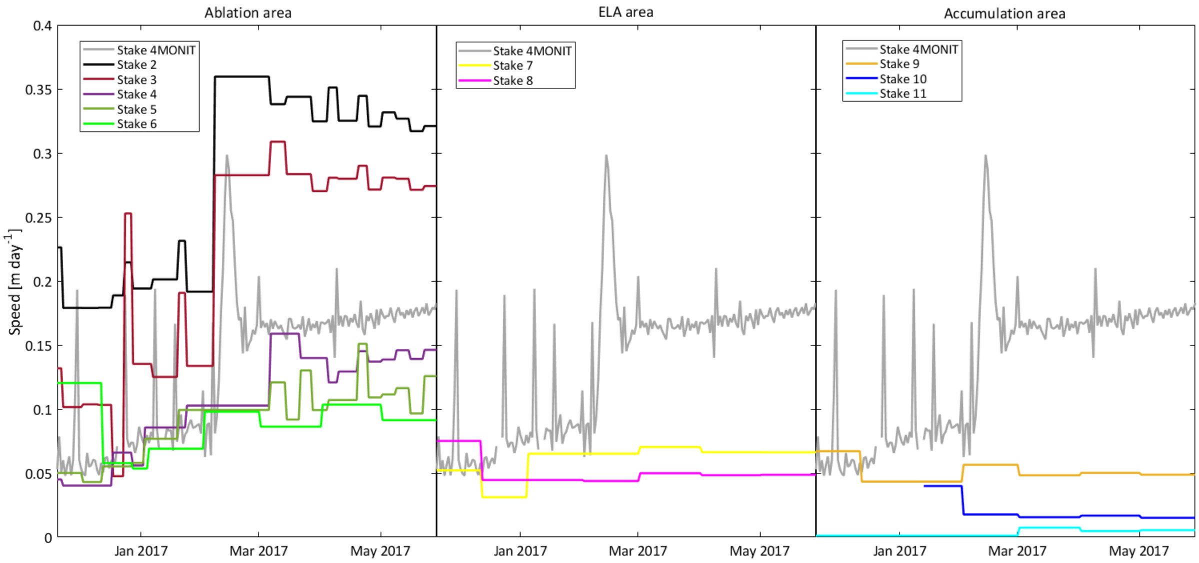

Figure 4. Time series of Hansbreen speed along its center line at stakes number 2 to 11 for three different glacier areas compare to glacier speed at 4MONIT stake.

Duval et al., 1983; Glen, 1955). It is likely that this closure explains the 10 m recorded (from the sensor at 28 m above the bed to the final ~38 m above the bed) upwelling observed at the end of February. After this upwelling ends in early March, the slow decrease in water level is due to water leaving this cave system. It is not likely that water is directly draining out the front of the glacier into the fjord, because a decrease in subglacial water volume would likely cause a decrease in ice velocity, not an increase as shown here. Therefore, the local decrease in water level is likely due to creep closure of the system, pushing water to the surrounding bed, and decreasing the effective pressure (Cowton et al., 2016; Hewitt, 2011; Werder et al., 2013), after which it may or may not leave the glacier. This warm winter event may have also influenced the velocity of the following 2017 summer, which had an average near-terminus velocity 18 % higher than the last 10 years (172 m.y^{-1} compare to 145 m y^{-1}).

If we assume stake 2 as representative of the front velocity (Figure 1), its winter baseline velocity (from 2016-12-01 to 2017-02-01) is ~0.19 m day^{-1}. After the warm winter event its average velocity is ca. 0.34 m day^{-1} until the end of the accumulation season (end of May). The velocity increase from this warm event, lasting more than 3 months, is ~0.15 m day^{-1}. From this, 80 % more ice entered the fjord in the 2016/2017 winter than if this 1-week warm event had not occurred, or ~10 % more ice compared to the annual average.

Data gaps and velocity spikes - In December we show two gaps in the temperature record (6 days and 5 days) during which precipitation events occur (Figure 3). There is no observed water level fluctuations (water level remained within 28 m of the bed) and no coincident velocity increase. However, there are 3 one-day-long velocity increases prior to the February event, all while temperature are below 0° C. The cause of these events is not clear, nor are they significant when compared to equally large but multi-month velocity increase later in the record.

Our dataset also highlights winter storage and discharge of water (e.g. Hodge (1974); Hodgkins (1997); Hodson et al. (2005); Jansson et al. (2003); Wadham et al. (2000)). Our observed water level remains more or less steady at 38 m above the bedrock with k values around 0.55 (Figure 3), providing clear evidence of multi-month storage of large volumes of water. However, water can move dynamically and discharge to the distributed system while appearing more or less steady at the location of the logger, if the subglacial system closes equal to the volume discharged. We note that Pitcher et al. (2020) attribute their winter glacier discharge to storage of summer runoff, but acknowledge a warm event 10 days prior to their observation. Our data suggests that warm events may fill the system, and there may not be enough evidence to properly attribute the source of the discharge observed by Pitcher et al. (2020).

Similarly, Vijay et al. (2019) identified "type-3" glaciers in Greenland which are characterized by winter speedup events associated with subglacial meltwater activity. They assign the meltwater to different sources: basal meltwater, ocean water infiltrating into the subglacial system, and meltwater that did not evacuate through channels during the melt season and was retained in the firn and ice body but did not look at potential winter warm events, of which there are an increasing number in Greenland.

Contrary to Łupikasza et al. (2019), we show here that during this warm event water does percolate through the winter snow pack into the IDS. This is likely both rain and meltwater. Therefore, winter rainfalls, which were previously estimated to contribute to 9 % of the winter accumulation of Hansbreen (Łupikasza et al., 2019) with no loss term, should not be considered as only a mass-gain term. Some portion of the rainfall should be taken as an indirect mass-loss parameter.

The observations here are not unique to this glacier or Svalbard. After a warm winter event in Iceland, a Glacsweb wireless probe installed at the Skálafellsjökull glacier bed by Hart et al. (2019) observed a similar water pressure pattern. After an initial water pressure increase due attributed to the warm winter event, they recorded a sharp water pressure decline followed by a slow rise on subsequent days until the next melt event (Hart et al., 2019). Other Arctic glaciers may also be susceptible to these events. As the climate warms, precipitation onto the Greenland ice sheet is likely to shift towards a higher fraction of rain in the total precipitation (Bintanja & Andry, 2017; Boisvert et al., 2018; Lenaerts et al., 2020; Screen & Simmonds, 2012). If glacier dynamics models do not take into account the increase in off-season rain shown by regional climate models, then their ability to model the dynamical changes may be compromised, with related limitations in their ability to properly estimate sea level rise.

6 Conclusions

We show an Arctic glacier, as a result of a single winter warm event, has its average winter velocity more than double and remain at more than double the baseline for the remainder of the winter. The velocity increase appears to be sustained by englacial and subglacial water storage, evident by flooding of an instrumented cave, two and a half days after the warm event. After the warm event and some re-arranging of the subglacial hydrology, water backs up into the englacial system from the subglacial system. Within 10 days of the event a nearly steady state is reached, albeit with a small decrease in water level and continued small increase in ice velocity. We attribute this to water transfer from the discrete system to the distributed at the base of the glacier.

Warm winter events in the Arctic are being reported more often, and predicted to occur more often in a warming climate. We show these warm events can lead to large and sustained increases in ice velocity. Arctic tidewater glaciers are currently the most significant contributor to eustatic sea level rise. Further studies linking the atmosphere, ice velocity, and the winter subglacial hydrologic system are needed to quantify this contribution to sea level rise.

Acknowledgments

Data Availability Statement

All the data are archived at the Polish Polar Data Base: `http://ppdb.us.edu.pl/geonetwork/srv/eng/catalog.search#/home`. Velocity data of 4MONIT stake and of the center line of Hansbreen are respectively available at `http://ppdb.us.edu.pl/geonetwork/srv/eng/catalog.search#/metadata/6e8f320d-4c06-40ce-86cc-f8561d3df4bb` and `http://ppdb.us.edu.pl/geonetwork/srv/eng/catalog.search#/metadata/8c5219c5-2adb-40a5-a9de-eedcf8d0c7da`. Air temperature and precipitation are respectively available at `http://ppdb.us.edu.pl/geonetwork/srv/eng/catalog.search#/metadata/e5e66a63-126d-49e1-bebe-c623becfb5d8` and `http://ppdb.us.edu.pl/geonetwork/srv/eng/catalog.search#/metadata/6603c86f-3194-4fbd-a7e8-5c0bbf430c94`. Water level data of C.C are available at `http://ppdb.us.edu.pl/geonetwork/srv/eng/catalog.search#/metadata/0a4570d3-576b-45e7-947e-737f610d976f`.

This work has been financed by the Centre for Polar Studies, University of Silesia – the Leading National Research Centre (KNOW) in Earth Sciences (2014–2018), No. 03/KNOW2/2014. Glaciological, hydrological and meteorological data were processed by the University of Silesia data repository within project Integrated Arctic Observing System (INTAROS). This project has received funding from the European Union's Horizon 2020 research and innovation programme under grant agreement no. 727890. We wish to thank colleagues from the Polish Polar Station at Hornsund and the Institute of Geophysics Polish Academy of Sciences for hospitality and logistic support during field missions. Access to the meteorological data from the Hornsund station and GNSS data provided by the Institute of Geophysics, Polish Academy of Sciences is kindly acknowledged. The study used equipment and logistical resources belonging to the Polar Laboratory of the University of Silesia in Katowice. Thanks to the VENDYS foundation for its support. Thanks to Barbara Barzycka for providing data on ELA. Léo Decaux acknowledges the field assistance of Anton Sedlák, Kacper Konior, Jakub Medrzycki, Andreas Alexander, Konrad Osienienko, Sabina Kucieba, Kamil Pałkowski, Mateusz Mandat, Mariusz Czarnul, Maciej Błaszkowski, Robert Pogorzelski, Dominika Dabrowska, Grzegorz Piotrowski, Benita Siepierska, Jakub Szewczyk, Bartosz Siek and Łukasz Pawłowski. The authors declare there are no real or perceived financial conflicts of interest, or other affiliations for any author that may be perceived as having a conflict of interest with respect to the results of this research.

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