Winter speed-up of quiescent surge-type glaciers in Yukon, Canada

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Abstract

Glacier surges often initiate in winter, but due to the normal summer speed-up, their mechanism remains unclear. To address this question, we used radar images to examine spatial-temporal changes in the ice velocity of surge-type glaciers near the border of Alaska and Yukon. We found significant upstream accelerations from fall to winter, regardless of surging episodes. Moreover, whereas the summer speed-up was observed downstream, the winter speed-up propagated from upstream to downstream. Given the absence of upstream surface meltwater input in winter, we suggest the presence of water storage near the base that do not directly connect to the surface yet can promote basal sliding through increased water pressure as winter occurs. Our findings have implications for modeling of glacial hydrology in winter, which may affect future glacier dynamics.

1 Introduction

Ice flow on mountain glaciers and ice sheet typically has its greatest acceleration from spring to early summer, followed by deceleration in mid-summer to fall (e.g., Iken and Bindschadler, 1986; Zwally et al., 2002; MacGregor et al., 2005; Bartholomaus et al., 2008; Sundal et al., 2011). These speed changes are attributed to subglacial slip associated with water pressure changes, and these changes arise from seasonal variability of meltwater input and the evolution of the subglacial hydraulic system (Schoof, 2010; Hewitt, 2013; Werder et al., 2013). From spring to early summer, meltwater from the surface reaches the bed, and develops an "inefficient" drainage system, in which water flow channels are not well developed, producing a high basal water pressure. The high water pressure increases basal slip, which increases the surface velocity. As the amount of meltwater increases, the basal drainage system becomes more and more "efficient" due to the enlarging channels (Röthlisberger, 1972). The larger channels allow a higher meltwater flux with lower water pressure that lead to a gradual decrease in the surface velocity. In late summer to fall, when the meltwater input terminates, the surface velocity has its yearly minimum. These factors influence surface ice speeds from spring to fall, but what factors control the ice speeds in winter?

Several studies reported surface ice speeds in winter to be in between the early summer maximum and early fall minimum (e.g., Iken and Truffer, 1997; Sundal et al., 2011; Burgess et al., 2013a). Some recent studies also indicate that the amount of surface meltwater in summer can influence the velocity evolution in winter, in a way that reduces the annual ice flow (Burgess et al., 2013b; Sole et al., 2013). Due to the harsh environment and logistic problems, there have been relatively few comprehensive velocity measurements throughout wintertime particularly in the middle-to-upstream regions of mountain glaciers.

Nevertheless, it is well-known that glacier surges often initiate in winter, exhibiting orders-of-magnitude speed-up and resulting in km-scale terminus advance (Meier and Post, 1969; Raymond, 1987). Both the wintertime surge initiation and the intermediate values of winter speed have been interpreted as being caused by cavity closure and the subsequent water pressure increase, starting with the surge mechanism proposed for the 1982-83 surge at the Variegated Glacier by Kamb et al. (1985). Even in winter, there may be some remnants of summer meltwater that can increase the water pressure. However, it remains an open question why and how the water pressure increase and subsequent speed-up can be maintained without further input of meltwater from the surface. Do the surface velocities monotonously increase from later summer to the next spring? Such an increase is often assumed, but the process would require some extra sources of water to maintain the higher water pressure. The wintertime dynamics of sub- and englacial water are thus yet to be fully understood. Reaching an understanding requires new continuous measurements.

The St. Elias Mountains near the border of Alaska, USA, and Yukon, Canada (Fig. 1) contain numerous surge-type glaciers (Meier and Post, 1969). But only a few of these have been studied and reported in the literature (e.g., Clarke et al., 1984; Truffer et al., 2000; Flowers et al., 2011; Burgess et al., 2012). Our understanding of surge-type glacier dynamics is still limited (Raymond, 1987; Harrison and Post, 2003; Cuffey and Paterson, 2010), because few detailed observations have been performed over a complete surge-cycle.

Recent advance in remote sensing techniques allow us to survey the ice-velocity distribution over the entire St. Elias Mountains. Here we present the spatial and temporal changes in the ice velocity for the surge-type glaciers there, focusing particularly on the seasonal cycle during the quiescent phases to better understand the wintertime behavior. Three glaciers (Chitina, Anderson, and Walsh) significantly accelerate in the upstream from fall to winter, with speeds that are comparable to, and sometimes higher than those in the next spring to early summer. This is apparently in contrast to previously observed winter velocities (e.g., Iken and Truffer, 1997; Sundal et al., 2011) that appeared to be significantly slower than the velocities in spring and early summer. We interpret these observations by speculating the presence of englacial water storage, and discuss its implications for the surge mechanisms.

Understanding the dynamics of surge-type glaciers is also important to better simulate future ice dynamics in St. Elias Mountains. Significant contributions of the Alaskan glaciers' retreat to the possible sea-level rise due to the global warming have been estimated (Radić and Hock, 2011), but projections of glacier mass balance assume non-surge type glaciers whose dynamics are only affected by long-term climate changes. Although the dynamics of surge-type glaciers itself is not directly related to the climate change, there have been several pieces of evidence for the impact of climate change on surge cycle (e.g., Harrison and Post, 2003; Frappé and Clarke, 2007).

2 Data sets and analysis method

2.1 ALOS/PALSAR data

We processed phased array-type L-band (wavelength 23.6 cm) synthetic aperture radar (PALSAR) images from the Advanced Land Observation Satellite (ALOS) operated by the Japan Aerospace Exploration Agency (JAXA). Data was acquired along multiple paths (Fig. 1, Table 1). ALOS was launched on January 2006, and its operation was terminated on May 2011. Thus, the datasets for the study area were acquired only from December 2006 to March 2011. The details of the datasets are listed in Table 1. Only the FBS (fine-beam single-polarization mode) and FBD (fine-beam dual-polarization mode) data are used in this study. We use Gamma software to process level 1.0 data to generate single look complex images

(Wegmüller and Werner, 1997) and run pixel-offset tracking analyses. See Table 1 for more detail of the datasets.

2.2 Pixel offset tracking

The pixel-offset tracking (or feature or speckle tracking) algorithms used in this study are based on maximizing the cross-correlation of intensity image patches. The method closely follows that used by Strozzi et al. (2002) and Yasuda and Furuya (2013). We used a search patch of 64 × 192 pixels (range × azimuth) with a sampling interval of 4 × 12 pixels. But, due to its larger size for Hubbard Glacier, we used a search patch of 128 × 384 pixels. We set 4.0 as the threshold of the signal-to-noise ratio and patches below this level were treated as missing data. The FBD data are oversampled in the range direction so that the range dimension is the same as that of the FBS data.

In the pixel-offset tracking, we corrected for a stereoscopic effect known as an artifact offset over rugged terrain (Strozzi et al., 2002). That is, because of the separation between satellite orbital paths, the effect of foreshortening also differs in the offsets. We reduced the artifact by applying an elevation-dependent correction, incorporating the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) global digital elevation model (GDEM) version 2 data with 30-m resolution. We applied the same method described by Kobayashi et al. (2009) and confirmed that there remained few topography-correlated artifact offsets.

Using both range and azimuth offset data, we derived the surface velocity data (Fig. 1) by assuming no vertical displacements. The studied glaciers are gently sloped at approximately 1-2 degrees, and thus, the vertical component is much smaller than the horizontal component. In addition, we derived the velocity map using image pairs that were temporally separated by at most 138 days. The glaciers' thinning during this period should be negligibly small in comparison to the horizontal movement of the glaciers. We averaged the velocity data over the ~350 × 350-m^2 area along the flow line and, from the standard deviation at each area, estimated the measurement error to be below 0.1 m/d.

Using two data images with ALOS/PALSAR's 46-day intervals acquired at non-deforming areas (Kobayashi et al., 2009), the uncertainties of offset tracking data in the rugged terrain have been estimated to be ~0.3-0.4 m. Assuming linear temporal evolution, the errors in the velocity estimate are inferred to be below 0.1 m/d.

3 Observation results

Although surging episodes occurred at Lowell, Tweedsmuir, and Ottawa, here we focus on winter speed-up signals at surge-type glaciers that were in their quiescent phase during the analysis period. These occurred at seven glaciers (Chitina, Anderson, Walsh, Logan, Hubbard, Agassiz, and Donjek). The Chitina, Anderson, Walsh, and Logan Glaciers, which are the major surge-type glaciers of the Chitina River valley system (Clarke and Holdsworth, 2002), could be examined with the highest temporal resolution because of the overlap of multiple satellite tracks. Major 17 glaciers in the region are shown in Figure 1.

Figure 2 shows flow velocity at Chitina Glacier from oldest at top left to most recent at bottom right. Notice that the flow velocity in the upstream gradually increases from fall to winter every year (Fig. 2c-f, g-j, k-o, u-z). Starting in fall 2009, the velocity increases at the confluence between Chitina and Ottawa Glacier (Fig. 2l). On Feb-Mar 2010, it speeds up to 4 m/d at Ottawa Glacier (Fig. 2p-q), which we regard as a glacier surge (see the supplementary material). At the same time, the velocity in the upstream region of Chitina Glacier gradually increases as winter approaches (Fig. 2k-o). In contrast to the surge, the winter speed-up occurs every winter, which thus indicates that the wintertime acceleration in the upstream of Chitina Glacier is independent of the surge at Ottawa Glacier. Moreover, the winter speed in the upstream region is comparable to and sometimes higher than that in spring/early summer in 2010 (Fig. 2s), which we believe had not been observed before. The higher speed in the middle to downstream (Fig. 2q-t) may have been triggered by the surge at Ottawa Glacier. Similarly high winter speeds were also detected at other surge-type glaciers.

Figure 3 shows the spatial-temporal evolution of ice velocity of four glaciers along their flow lines. At Chitina Glacier, the winter velocities in the upstream region exceed 0.5 m/d, which is significantly greater than the fall velocities of ~0.3 m/d regardless of the surge signal at Ottawa glacier (Fig. 2l-t, Fig. 3a). At the 20-km point upstream on Anderson Glacier (Fig. 3b), the winter speed is more than double the fall speed. Along the upstream segment on Walsh Glacier (Fig. 3c), the winter speed is more than 50% greater than the fall speed.

Consider the distinction between upstream and downstream seasonal trends. Although the downstream speeds in early summer are faster than those in winter, the upstream speeds in winter are comparable to, and sometimes faster than those in early summer. For instance, at the 20-km point upstream on Anderson Glacier, the velocity is ~0.5 m/d in early summer 2010 but exceeds 0.7 m/d in winter of 2009/2010 and 2010/2011. Similarly, at 20-25 km upstream on Walsh Glacier, the velocity is 0.3-0.5 m/d in early summer 2010 but 0.6-0.8 m/d in winter. Moreover, in contrast to the upglacier propagation of summer speed-up observed in the ablation zone of glaciers in Greenland (Bartholomew et al., 2010), here the higher-velocity area expands from upstream in fall to downstream in winter. This downglacier propagation is clearest at Anderson Glacier (Fig. 3b). These trends apply to longer glaciers as well. Logan Glacier, with nearly twice the length of the above three glaciers, has a broad segment in the middle that accelerates from fall to winter (Fig. 3d). In addition, the winter velocities appear to increase from one year to the next, indicating the initiation of a new surging episode (Fig. 3d).

Although we could not obtain quality summer velocity data for each year due to large intensity changes associated with surface melting, the glacier dynamics at lower reaches is consistent with previous findings. For example, Figure 3 shows summer speed-up signals in 2010 in the lower to middle reaches of each glacier. In addition, compared to the gradual downglacier propagation of the winter speed-up noted above, the summer speed-up in the lower reaches appears to occur primarily over a shorter period.

For Hubbard Glacier, the only tidewater glacier in the study area, the ~15 km-length section in the midstream region has velocities in January and February that are ~33-60% greater than the velocities of the previous August to October (Figs. 4a, d, e, and h). The significant speed-up during the 2009 winter is most likely associated with a small surge in the upper tributary (Fig. 4e). The much smaller tributary in the upper reach of Malaspina Glacier (Fig. 1) also exhibits greater velocities in winter, as does Agassiz and Donjek Glacier (Fig. 1, Fig. 5), suggesting that the winter speed-up mechanism is independent of the glacier’s size.

Consider Agassiz and Donjek Glacier. At Agassiz Glacier, the winter midstream speed-up and downglacier propagation occur from fall to winter in the 2007-2008, 2009-2010, and 2010-2011 seasons (Fig. 5a). Moreover, the winter velocities in 2008 and 2011 are clearly greater than the fall velocities in the corresponding years. The greater velocities in the summer 2010 indicate a summer speed-up. The greatest seasonal fluctuations occur near 10 km, outlined in black in the figure. At Donjek Glacier, the black-squared segment mid-glacier (Fig. 5b) shows winter velocities that are greater than the fall velocities. However, the downglacier propagation is not clear in the Donjek case.

4 Discussion

According to the average air temperature at Yakutat Airport provided by The Alaska Climate Research Center data (http://akclimate.org), the monthly average temperature from 2006-2011 is about 0.2 ℃ in November, and about -2 ℃ for December, January, and February. Almost all of our study area is above 1000 m a.s.l., except Agassiz Glacier, which extends from 450 to 1100 m a.s.l. Thus, the wintertime temperature is significantly below freezing, so there should be little surface meltwater during winter. Moreover, each glacier's location in this study is much higher than that at Variegated Glacier, which is a temperate glacier. Under such circumstances, it is likely that the mechanisms of winter speed-up and its downglacier propagation are different from those of the summer speed-up that usually propagates upglacier. Also, the detected annual winter speed-up in the upstream is up to 100% too high to be explained by snow accumulation.

The observed winter speed-up in the upstream region may be regarded as a "mini-surge" (Humphrey and Raymond, 1994). However, not all previously reported mini-surges occurred in winter. For instance, the mini-surges prior to the 1982-1983 surge at Variegated Glacier occurred in summer (Kamb et al., 1985; Kamb and Engelhardt, 1987). A mini-surge defined in Kamb and Engelhardt's paper indicates dramatically accelerated motion for a roughly 1-day period, which occurred repeatedly during June and July in 1978-81. Although Kamb et al. (1985) noted an anomalous increase in wintertime velocities since 1978, the measurements were done only once in September and once in June (Raymond and Harrison, 1988), and thus they may include the spring speed-up signals as pointed by Harrison and Post (2003). No comprehensive wintertime velocity observations have been done upstream.

We now compare our findings to previous studies. Iken and Truffer (1997) found a gradual speed-up from fall to winter at the ~2-km-long downstream section of the temperate Findelengletcher in Switzerland, where the speed continues to increase, reaching as maximum in summer. In contrast, our observed winter speed-up occurs in the upstream region, and speed does not continue to increase after winter. Sundal et al. (2011) examined how ice speed-up and meltwater runoff are related at land-terminating glaciers in Greenland. The ice speed-up is affected by the amount of surface runoff each year, which differs between high and low melting years. The results indicate that the ice speed in a high melting year gradually increases from fall to winter. However, the ice speed does not accelerate in low melting years. Moreover, they did not report the spatial distribution of speed during winter, and the maximum speed is apparently observed in early spring to summer. Our velocity data do not simply indicate the gradual speed-up from fall to next spring. The winter speed-up initiates upstream, and the maximum speed in winter is comparable to that in early summer. As some of the glaciers could not be examined with a high temporal resolution, it is likely that there are other winter speed-up glaciers.

How can we explain the observed winter speed-up signals? First, we argue that the mechanism proposed by Kamb et al. (1985) for the Variegated Glacier does not apply here. In that mechanism, the efficient tunnel-shaped drainage system, which is present in summer, may provide a less efficient distributed system in early winter due to depletion of surface meltwater and the destruction of conduits by creep closure. Thus, the subglacial water pressure may greatly increase. For our observed winter speed-up to be explained by this mechanism, there would have to be an efficient drainage system. Although such an efficient drainage system is often observed near the terminus (Raymond et al., 1995; Werder et al., 2013), the winter speed-up is observed upstream, far from the terminus. In addition, even if there exists meltwater remnant in the upstream region, it is unclear how the subsequent speed-up can be maintained without further input of meltwater from the surface. Thus, we need to consider a mechanism that can trap water in the upstream in winter so that the subglacial water pressure can be maintained high enough to generate basal slip.

One such mechanism was proposed by Lingle and Fatland (2003). In that study, using the few ERS1/2 tandem radar interferometry data with the 1-3 day's observation interval, they similarly detected a faster speed in winter than in fall at the non-surging Seward Glacier in the St. Elias Mountains. They also found localized circular motion anomalies at both surging and non-surging glaciers that indicated local uplifting and/or subsidence caused by transient subglacial hydraulic phenomena. Combining their observations with earlier glacier hydrological studies, they proposed a model of englacial water storage and gravity-driven water flow toward the bed in winter that applies to both surge-type and not surge-type glaciers.

Few winter speed-up observations have been made since Lingle and Fatland (2003), but our data suggests that winter speed-up may not be a rare phenomenon. Each local uplift and/or subsidence event in the Lingle and Fatland study must be a transient short-term process, episodically ocurring in places. We could not observe such localized signals in our offset-tracking displacements because our observation interval, at least 46 days, is much longer than the 1-3 days in Lingle and Fatland (2003). Nevertheless, we propose that both Lingle and Fatland's and our observations are caused by the same physical processes. This is because the locally increased basal water pressure could increase basal sliding and contribute to larger horizontal displacements.

Till deformation is another mechanism to cause glacier surge (e.g., Cuffey and Paterson, 2010), and some glaciers in Alaska and Yukon have till layers. For example, Truffer et al. (2000) examined surface velocity and basal motion at ice-till interface at Black Rapid Glacier in the Alaska Range, finding that the large-scale mobilization of subglacial sediments plays a dominant role in the surge mechanism. However, based on Coulomb-plastic rheology for the till deformation (e.g., Clarke, 2005), substantial till deformation requires a high basal water pressure. So, regardless of the presence of till layer, the mechanism for winter speed-up should include a process in which a high basal water pressure can be kept during wintertime.

Schoof et al. (2014) recently reported wintertime water pressure oscillations at a surge-type glacier in Yukon, and interpreted them as spontaneous oscillations driven by water input from englacial sources or ground-water flow. But without flow velocity data, they could not correlate the wintertime drainage phenomenon to glacial dynamics. The present observations though are consistent with the englacial water storage model of Lingle and Fatland, and thus may help explain our observed upstream glacier speed-ups in winter.

Although the englacial water storage model may explain the winter speed-up, the specific water-storage system remains unknown (Fountain and Walder, 1998). One plausible form of englacial water storage is the basal crevasses observed by Harper et al. (2010) at Bench Glacier, Alaska. Such crevasses have no direct route to the surface, yet can store significant volumes of water near the bed. Thus, water in the basal crevasses may generate high pressure when they become constricted due to creep closure in winter.

The formation of basal crevasses in grounded glaciers requires a high basal-water pressure that may approach the ice overburden pressure and/or longitudinally extending ice flow (van deer Veen, 1998). Although such crevasses have not been detected in this area, their restrictive conditions might explain our observations of uncommon winter speed-up signals and the distribution of surge-type glaciers in the area.

5 Conclusions

In this study, we applied offset tracking to ALOS/PALSAR data on glaciers near the border of Alaska and Yukon to show their spatial and temporal velocity changes in 2006-2011. Surging episodes occurred at three glaciers (Lowell, Tweedsmuir and Ottawa). For many of the quiescent surge-type glaciers around the St. Elias Mountains, upstream accelerations occurred from fall to winter and then propagated downstream. The winter speeds in the upstream regions were comparable to, and sometimes faster than those in spring to summer. Combining the absence of upstream surface meltwater input in winter with insights from some previous studies, we speculate that sizable water storage may be present near the bottom of glaciers, not directly connected to the surface, yet can enhance basal sliding by increased water pressure as they constrict in winter. Further observational and theoretical studies are necessary to decipher the winter speed-up mechanisms and determine if such water storage systems exist.

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Table 1. Data list of the ALOS/PALSAR.

Sensor/Path	Frame	Master	Slave	Mode	[#]Bperp (m)	Span (day)
PALSAR/241	1190–1210	20070829	20071014	FBD–FBD	597	46
		20080114	20080229	FBS–FBS	796	46
		20090116	20090303	FBS–FBS	529	46
		20100119	20100306	FBS–FBS	756	46
		20100306	20100421	FBS–FBS	353	46
		20100421	20100606	FBS–FBD	104	46
		20100606	20100722	FBD–FBD	122	46
		20100722	20100906	FBD–FBD	332	46
PALSAR/243	1200 –1220	20061230	20070214	FBS–FBS	1342	46
		20070817	20071002	FBD–FBD	425	46
		20071002	20080102	FBD–FBS	627	92
		20080102	20080217	FBS–FBS	1041	46
		20080819	20090104	FBD–FBS	1779	138
		20090104	20090219	FBS–FBS	652	46
		20090822	20091007	FBD–FBD	566	46
		20091007	20100107	FBD–FBS	726	92
		20100107	20100222	FBS–FBS	794	46
		20100825	20101010	FBD–FBD	505	46
PALSAR/244	1200–1220	20070116	20070303	FBS–FBS	1554	46
		20070903	20071019	FBD–FBD	474	46

continued

		20071019	20080119	FBD-FBS	799	92
		20080905	20081021	FBD-FBD	672	46
		20081021	20090121	FBD-FBS	874	92
		20090908	20091024	FBD-FBD	419	46
		20091024	20100124	FBD-FBS	960	92
		20100124	20100311	FBS-FBS	722	46
		20100911	20101027	FBD-FBD	504	46
		20101027	20110127	FBD-FBS	997	92
		20110127	20110314	FBS-FBS	840	46
PALSAR/245	1200-1220	20070920	20071105	FBD-FBS	655	46
		20071105	20071221	FBS-FBS	86	46
		20071221	20080205	FBS-FBS	884	46
		20080807	20080922	FBD-FBD	1027	46
		20080922	20081223	FBD-FBS	596	92
		20090810	20090925	FBD-FBD	671	46
		20090925	20091226	FBD-FBS	776	92
		20091226	20100210	FBS-FBS	690	46
		20100210	20100328	FBS-FBS	532	46
		20100328	20100513	FBS-FBD	169	46
		20100513	20100628	FBD-FBD	122	46
		20100628	20100813	FBD-FBD	486	46
		20100813	20100928	FBD-FBD	470	46
		20100928	20101229	FBD-FBS	614	92
		20101229	20110213	FBS-FBS	790	46

Bperp stands for the orbit separation distance perpendicular to the radar line of sight.

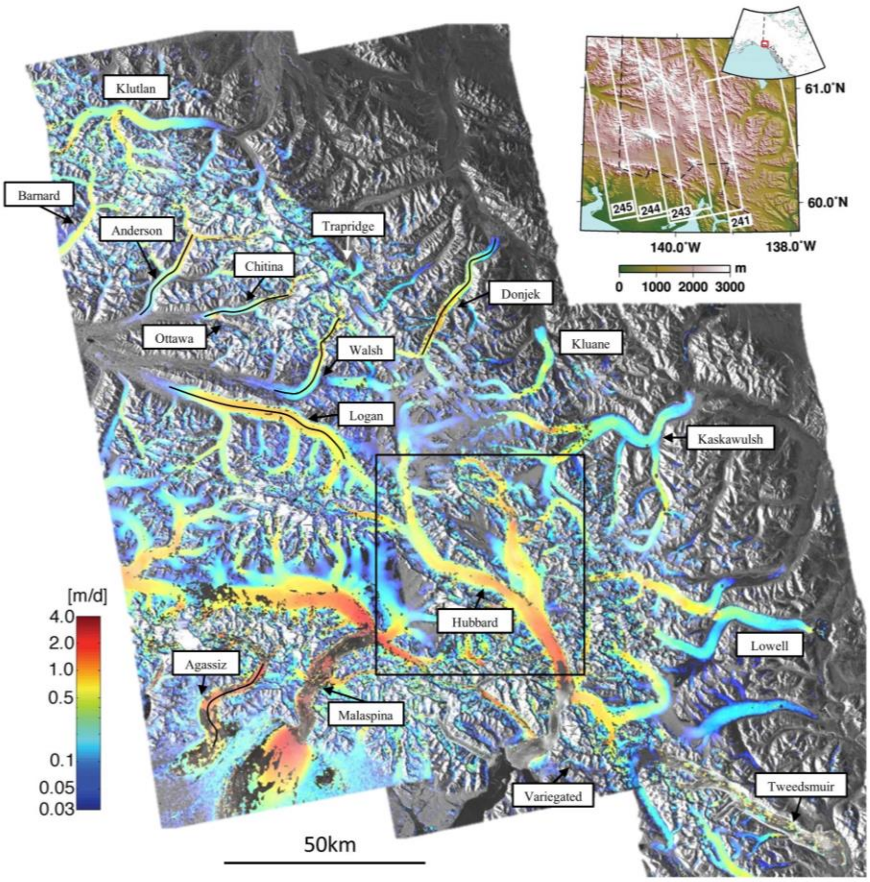

Figure 1. Composite ice-speed map of the study area. The individual maps for the study area were derived by intensity tracking between two PALSAR images. The left, middle and right velocity maps are derived from images pairs from 10 February 2010 and 28 March 2010 of Path 245, 30 December 2006 and 14 February 2007 of Path 243, 14 January 2008 and 29 February 2008 of Path 241, respectively. The square region around Hubbard Glacier is shown in Fig. 4. Black lines in some glaciers show the flow line. The upper right panel indicates the location and topography of the study area as well as the satellite's imaging areas.

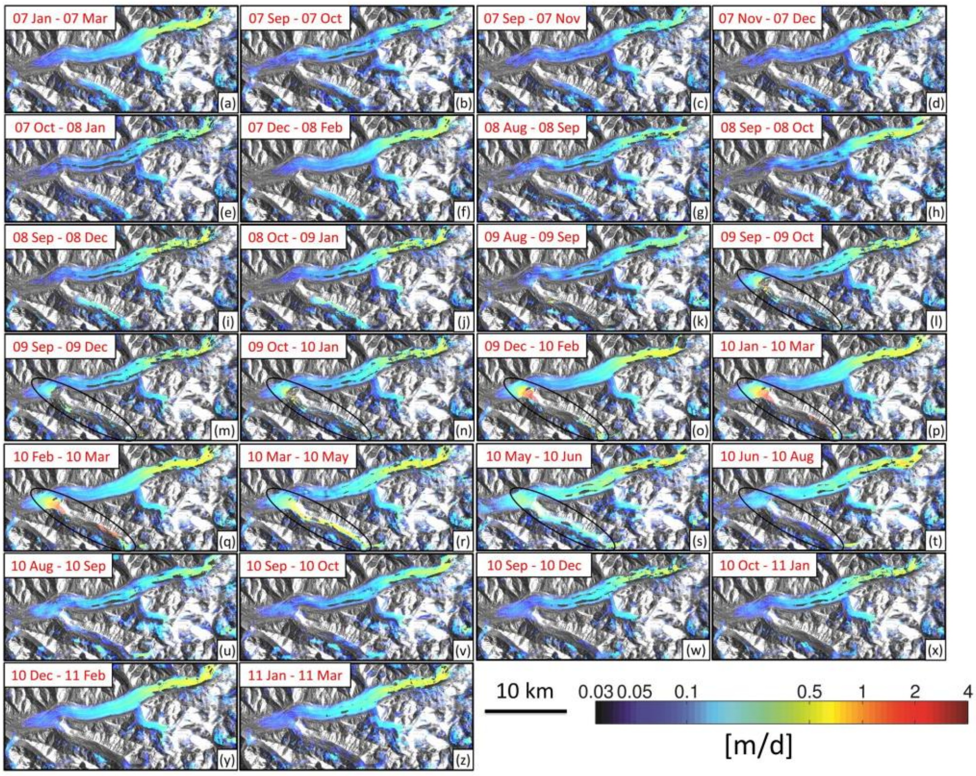

Figure 2. Surface velocity time-series (from upper left to lower right) at Chitina Glacier. Images are arranged in the order of middle date between the first and second acquisitions for each pair. The color scale is logarithmic. The black ovals mark a surge from autumn 2009 to summer 2010 on Ottawa Glacier. Details of the surge are in the supplementary material.

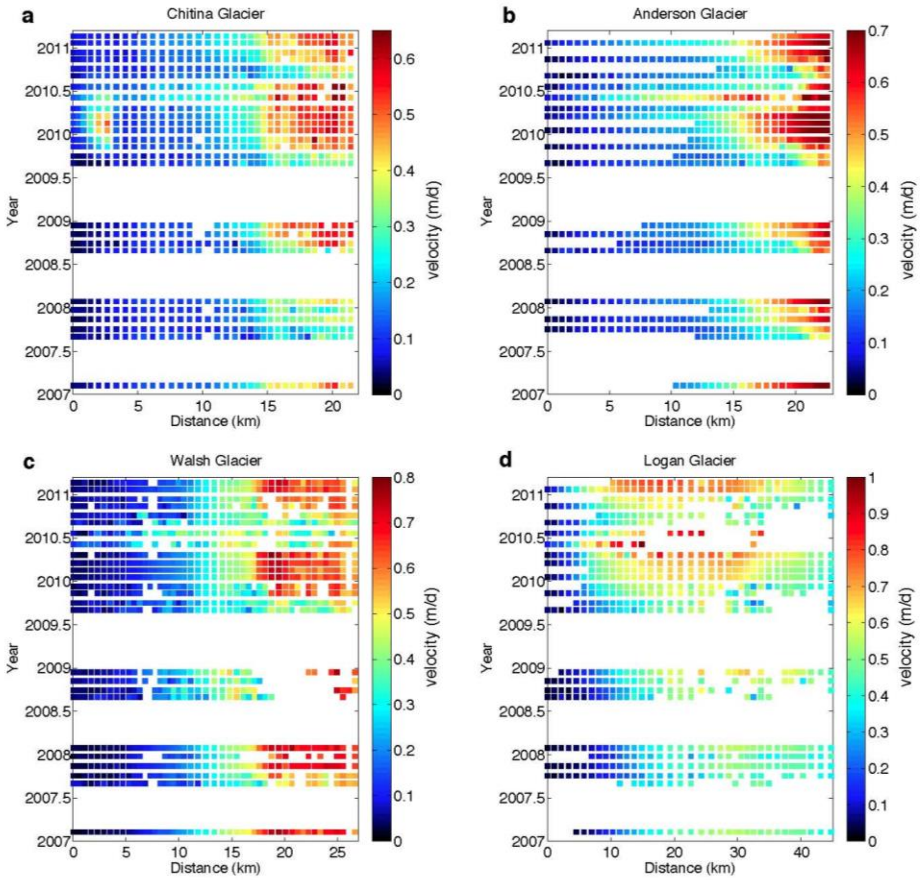

Figure 3. Time evolution of ice velocity profiles along the flow lines of Chitina, Anderson, Walsh, and Logan Glaciers. The flow lines are marked in Fig. 1.

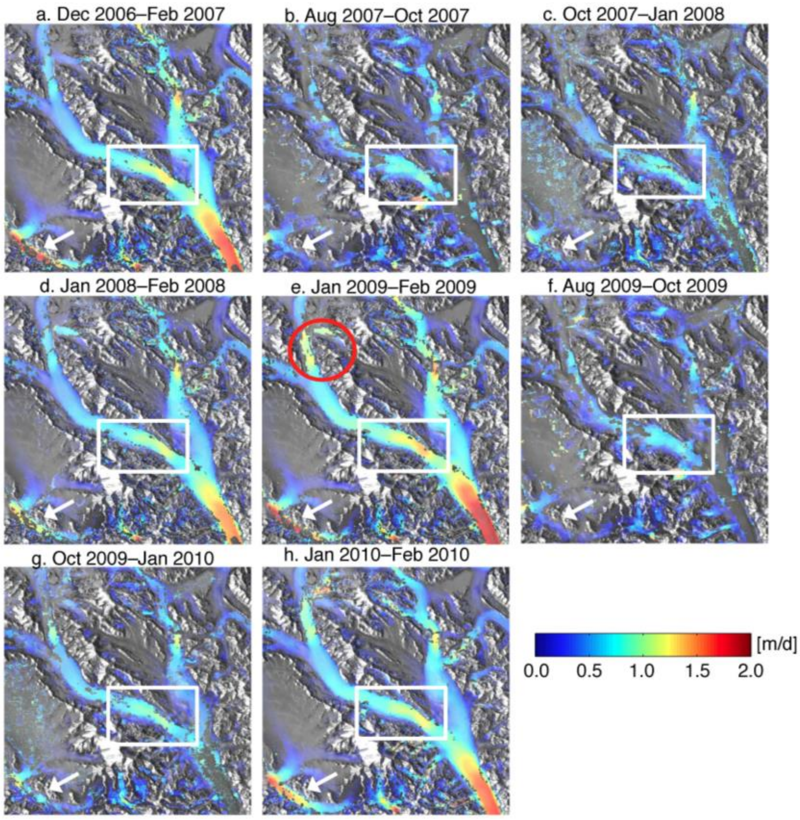

Figure 4. Spatial-temporal evolution of ice velocity at Hubbard Glacier and an upper tributary of Malaspina Glacier. The white square marks a region in which the velocity in winter (a, d, e, h) exceeds that of late summer and fall (b, c, f, g). The red circle in (e) marks a "mini-surge-like" signal in the upstream region during January-February 2009. The white arrow in that image shows a winter speed-up of an upper tributary of Malaspina Glacier.

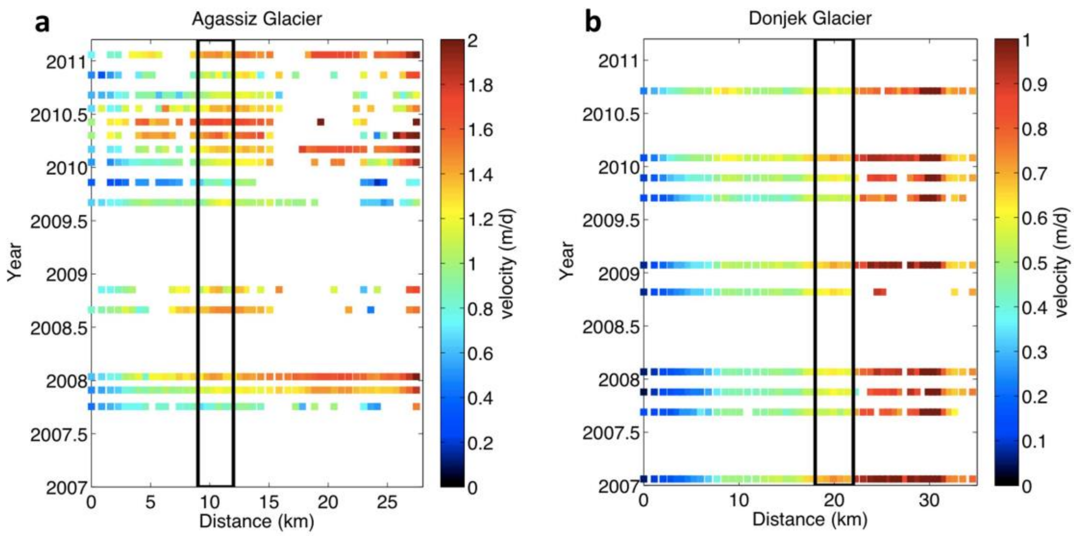

Figure 5. Temporal evolution of ice velocity profiles along the flow lines of Agassiz and Donjek Glaciers. The black box indicates the section showing clear seasonal changes.