Spatio-temporal velocity variability of Eastern Karakoram Glaciers observed by ALOS-1/2 data

（**ALOS1/2** データで観測された東カラコルム山脈の氷河流動速度の時空間変動）

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Dedicated to my parents, who broke the family traditions, left their family and farming life and migrated to a city to educate their kids, in the early 1980s. Due to their struggle and support, I am going to have the honor being a Ph.D. graduate from one of the leading university of the world

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Abstract

Unlike in most other regions, Karakoram glaciers are either stable or advancing, a phenomenon known as the Karakoram anomaly. Despite studies of glacier surges and the derivation of surface velocity maps, the spatio-temporal variability of glacier dynamics still remains poorly understood, particularly in the Eastern Karakoram Range. We use Advanced Land Observing Satellite/The Phased Array type L-band Synthetic Aperture Radar (ALOS/PALSAR)-1/2 data from 2007–2011 and 2014–2015 to examine detailed surface velocity patterns of the Siachen, Baltoro, Kundos, Singkhu, and Gasherbrum glaciers. The first three glaciers show considerable velocity variability (20–350 m a^{-1}), with clear seasonal patterns. Although all glaciers, except for Baltoro, flow slowest in 2015, the velocity structures are individual and vary in space and time. In Gasherbrum Glacier, peak surge-phase velocities are seasonally modulated, with maxima in summers 2006 and 2007, suggesting surface melt plays an important role in maintaining the active phase. Given the relatively close proximity of these glaciers we assume that surface melt timing and rates are comparable. We therefore argue that the observed spatio-temporal and inter-annual velocity patterns are determined by local and internal mechanisms, including englacial and subglacial hydrology, thermal processes and tributary configuration of each individual glacier.

1. INTRODUCTION

1.1. Why we study the glacial dynamics

High-mountain regions with glaciers act as ''water towers'' for the lowlands, as they provide meltwater not only to the people living close to the mountains, but also to the lowlands via runoff that recharges the river-fed aquifers. Also, through this runoff, such regions affect global sea level (Bolch and others, 2012). Pakistan is the 6th largest country of the world by population and its economy is mainly controlled by agriculture. The country is the 4th largest cotton producer of the world (Cotton Australia, 2016). The water is vital for agriculture. There are three major sources of water: 1) summer meltwater of winter snow, which is mostly accumulated in the northern Pakistan 2) monsoonal rainfall 3) underground aquifers. A yearly balanced cycle among all these resources is essential to maintain to a continuous and sustainable supply of water. Recently, the changing climate is not only influencing monsoonal patterns but also the glaciers' size. Therefore, a precise knowledge about the state of glaciers would be indispensable not only to understand the glacial dynamics in this region but will also to help policy makers have an insight about the condition of this important water resource.

1.2. Glacial dynamics and its significance

The glacial movement is the combination of:

i. Viscous flow (mechanical properties and stresses acting on the ice).

ii. Flow caused by the basal sliding, which is controlled by basal water pressure i.e. resulting from the contact between the glacier and the bed (governed by the glacial hydraulic system).

The viscous flow velocity is relatively small in the magnitude and is controled by the Glen parameter (Cuffey and Patterson, 2010), which can vary from the temperature changes in the space and time across the glacial length. On the other hand, basal sliding results from the basal water pressure, which-depends upon following factors:

i. Amount of melt water present in each season

   The amount of meltwater is mainly governed by three major factors: temperature, rainfall and amount of snowfall in the preceding winter. Any unusual increase in any of these factors can enhance the amount of available water that can infiltrate down to the base and can increase the basal water pressure to cause a hike in the glacial velocity.

ii. Efficiency of the hydraulic system

   Efficiency of the hydraulic system: One can divide the glacial hydraulic system into two broad categories: englacial and subglacial. And for the basal sliding, sub-glacial hydraulic system is the most important.

   Inefficient sub-glacial hydraulic system = high basal water pressure (high speed event)

   Efficient sub-glacial hydraulic system= low basal water pressure (low speed event)

iii Geothermal heat, Pressure melting by overburden ice.

Through the offset tracking of satellite images, one can measure the surface velocity of a glacier. The glacial surface velocities can be broadly divided into three categories:

0. No temporal velocity changes

a. Normal seasonal velocities (summers are usually faster than the winters).

b. Unusual high speed epochs (in which the glacial velocities show a sudden hike in the velocity. However, the sudden speed up may last for few months and is usually shorter than a year).

c. Surge events; these are high speed up events in which the velocities may increase by the factor of 10-100 and they sustain usually from more than a year to several years.

A surge-type glacier oscilated between a quiescent phase and an active phase, also known as the surging phase . During the active phase, the velocities are higher by a factor of 10-100 over a time period ranging from few months to years. In this phase, the terminal position may advance by up to several kilometers. On the other hand, the quiescent phase, also known as normal flow, can last tens to hundreds of years (Jiskoot, 2011). To explain differences in surge behaviour, Murray and others (2003) proposed the Alaskan-type hydrological regulation and Svalbard-type thermal-regulation mechanisms for the generation of high-pressure basal water that could drive an active surge. In a thermally-controlled surge, the initiation and termination phase lasts for several years before and after the peak phase, respectively, with all phases being independent of the season (Jiskoot, 2011). Jiskoot (2011) proposed the following steps for the thermally controlled surge:

i. Accumulation area increases in mass gradually and reaches the pressure melting point.

ii. Due to pressure melting the pore pressure increases which cause the deformation and dilatation of the underlying till thus enhancing the water storage capacity.

iii. Positive feedback: Enhanced deformation→ Frictional heat→ Further Melting→ Rising of pore-pressure

iv. This process continues until full surge development.

v. After the mass movement there will thinning of the accumulation area and due to the unavailability of enough mass to cause pressure melting, it will result in refreezing.

On the other hand, hydrologically-driven surges are characterized by a rapid acceleration and deceleration, and have tendency to initiate in winter (Raymond, 1987; Harrison and Post, 2003) and terminate in summer (Björnsson, 1998). Jiskoot (2011) proposed that hyrologically driven surge is initiated when there is switch from conduit dominated to distributed linked cavity system. The major reason is the collapse of conduit mainly due to: it may outgrow itself or the over burden pressure may cause the collapse. The termination of the surge will be when there is no sufficient surface melt to derive the surge.

The two models are also entirely different in terms of the origin of basal water. In the hydrological model, the water mainly originates from surface melt, while in the thermal-regulation model it is generated by pressure melting and frictional heating by the accelerated basal motion without input of surface meltwater; or melting by geothermal heat (←icecap in iceland).

In the current climate change scenario, where there are many reports of glacial retreat, the surge phenomenon gives an indication of anomalous glacial dynamic behavior in that region. It opens another corridor of research that what can be the possible mechanisms for such anomalous behavior.

1.3. Anomalous behavior of Karakoram Range Glaciers

The northern Pakistan lies at the junction of world's three highest, glacial covered mountain ranges (i.e. Hindukush, Karakorum and Himalaya). Due to the current climate changes, there are numerous reports that glaciers in Hindukush and Himalaya are receding (in phase with current climate change and resulting global glacial response). On the other hand, glaciers in Karakorum are either stagnant or increasing which is a phenomenon known as "Karakorum Anomaly" (out of phase with current climate change and resulting global glacial response).

The Karakoram is the fourth largest area of glacier cover on Earth (Dyurgerov and Meier, 2005), with an estimated glacial area of 18,000 square kilometers (Bolch and others, 2012). This region has a diverse vertical distribution of glaciers, ranging in elevation from 2300 meters a.s.l. in the Hunza valley to 6300 meters on K2 (Hewitt, 2006).

The Karakoram glaciers are fed by precipitation and avalanche. This precipitation has been increasing at elevations near 2500 and 4800 meters, but has maximum values between 5000 and 6000 meters (Hewitt, 2005). Within the region, westerly disturbances in winter bring around two-thirds of the high-altitude snow (Hewitt and others, 1988) and Indian monsoons in summer bring much of the remaining accumulation (Benn and Owen, 1998; Hewitt, 2005; Bookhagen and Burbank, 2010; Palazzi and others, 2015).

Concerning the terminus positions of these glaciers, mapping and field observations found a 5% retreat in the early twentieth century (Hewitt, 2011). The retreat slowed down in the 1970s, and then stabilized in the 1990s, when those in the high Karakoram starting to advance (Mayewski Jeschke, 1979; Hewitt, 2005). The trend remains variable, with the GRACE satellite gravimetric observations in 2003–2009 showing a net loss in mass of glaciers across the high Asian mountains, but the northwestern part including the Karakoram mountain range showing a gain in mass (Matsuo and Heki, 2010). Similarly, digital elevation model (DEM) data acquired from the Shuttle Radar Topographic Mission (SRTM) (Farr and others, 2007) and Satellite Pour l'Observation de la Terre (SPOT5) optical stereo imagery show that a slight gain in the mass in the central Karakoram glaciers occurred in the period 1999–2008 epoch (Gardelle and others, 2012).

The Karakoram anomaly may be caused by a higher transport of summer monsoonal moisture that not only increases snowfall at high altitudes, but also increases cloudiness and cooling in summers (Farhan and others, 2015; Zafar and others, 2016). Another view is that the reason for the resiliency of Karakoram glaciers could be by these due to a higher contribution of winter snowfall compared to monsoon snowfall (Kapnick and others, 2014).

But, to better understand the mechanisms of the Karakoram anomaly, we should consider that the Karakoram Range has a relatively high density of surge-type glaciers (e.g., Hewitt, 1969, 2007, 2011; Bhambri and others, 2017). Surge dynamics in the Karakoram have thus been extensively studied (Mayer and others, 2011; Quincey and others, 2011; Heid and Kääb, 2012; Rankl and others, 2014; Quincey and Luckman, 2014; Quincey and others, 2015; Paul and others, 2017). But it is still not clear whether the Karakoram anomaly has a correlation with the surge dynamics in the Karakoram Range. To answer this question, Copland and others (2011) suggested that the recent positive mass balance in the Karakoram played a role in the doubling in number of new surging episodes during the 14-year-period before and after 1990. However, the number of surge-type glaciers in their active phase has decreased since 1999, which is hard to explain only by the changes in mass balance, thus suggesting complexity in the surging mechanisms (Rankl and others, 2014). In particular, spatial and temporal changes in the thermal regime and the role of meltwater in en- and sub-glacial hydrology are poorly known for the Karakoram Range. Quincey and others (2015) concluded that neither the thermal, nor the hydrolological model of surging can account for the observed surges.

1.4. Purpose of the current research

To better understand the Karakoram Anamoly, the comprehensive in-situ observations would be ideal, but because of the remoteness and logistic issues in the Karakoram, it would be practically impossible to perform extensive in-situ observations, in the current study area. Instead, surface velocity maps over the Karakoram Range have been presented (e.g., Rankl and others, 2014). However, most previous velocity maps are either snapshots of a particular period or they have limited temporal resolution. Hence, the variability in seasonal velocity changes and even the presence of the seasonality itself have been uncertain.

To help clarify the mechanisms associated with glacial dynamics in the Eastern Karakoram Range, I report here on the spatial and temporal variability of glacier velocities with unprecedented resolution in this region. I use the L-band synthetic aperture radar (SAR) images acquired by the PALSAR sensors of ALOS-1 and ALOS-2, which can retain higher coherency due to its relatively long wavelength (23.8 cm) and allows us to examine detailed spatio–temporal changes in glacier velocities. The observed velocity variability will also be used as a reference dataset to diagnose if any episodes such as active surging and/or pulse event are ongoing.

1.5. Introduction of the study area

I focus on five glaciers in the Eastern Karakoram Range: the Baltoro, Siachen, Kundos, Gasherbrum, and Singkhu glaciers, as they have shown seasonal and inter-annual velocity variability in ALOS-1/2 data (Fig. 1). The relatively large size of these glaciers allows us to derive a detailed surface velocity distribution under the available spatial resolution of ALOS-1/2 images.

With a length of ~58 km, the Baltoro Glacier is one of the longest glaciers in the Karakoram Range (Copland and others, 2009). It is mostly debris-free in its upper reaches, but its debris cover becomes thick and extensive near its terminus (Quincey and others, 2009). Using ASTER imagery and GPS data, the lower-most 13 kilometers of the Baltoro Glacier were also examined for surface velocity (Copland and others, 2009).

The Siachen Glacier is ~72 km, the longest in the study area. Rankl and others (2014) have derived velocity data over Siachen Glacier, but a detailed spatial and temporal evolution of the glacier's velocity has remained elusive. Using multi-satellite images, Agarwal and others (2017) examined changes in its area, elevation, mass budget, and velocity. The velocity data were, however, limited to one summer pair in 2007 and one winter pair from December 2008 to January 2009, with detailed temporal changes being uncertain.

The Kundos Glacier has two tributaries, with the eastern being ~5 km longer than the western (Figs. 1 and 4c). Whereas the Kundos Glacier surface velocity was mapped by Rankl and others (2014), only one snapshot was given, and its spatio–temporal evolution remains uncertain.

For the Gasherbrum Glacier, a surge has been reported (Mayer and others, 2011, Quincey and others, 2011). I derive the surface velocity evolution during the transition period from the active phase to the quiescent phase, and complement previous findings on the surge dynamics.

The Singkhu Glacier is smaller than the Baltoro, Siachen, and Kudos Glaciers, and is located at altitudes between 4500–5800 m a.s.l.. Its seasonal and interannual velocity variability has not been reported before.

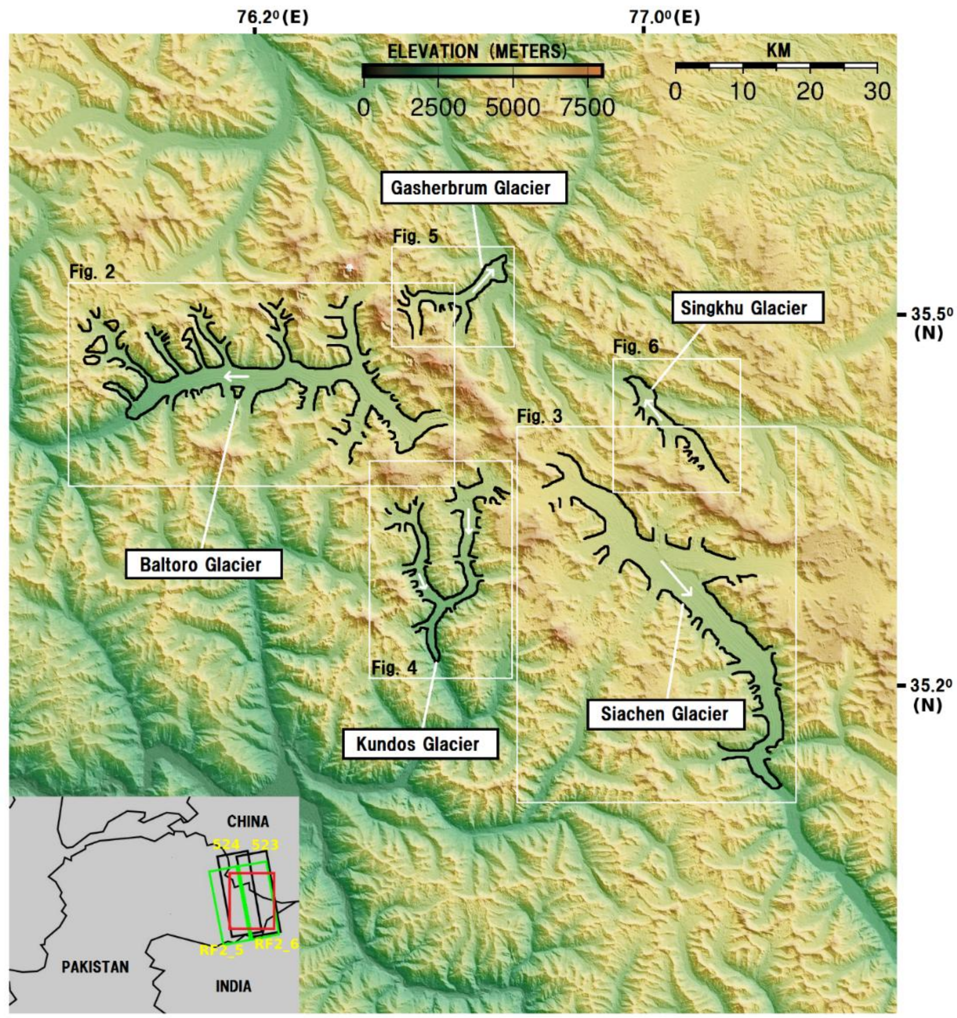


Figure 1. Location of the studied glaciers. Spatial-temporal evolution of surface velocities at each glacier is shown separately in Figures 2-6 and in supplementary materials file (Figures S1-S5). The spatial extent of the overview maps provided in Figures 2-5 are indicated by white rectangles. The ASTER-GDEM was used as a source of elevation in the background. Inset: Black and green rectangles show ALOS-1/PALSAR-1 observation paths (523,524) and ALOS-2/PALSAR-2 Look ids (RF2_5=RLF2_5, RF2_6=RLF2_6), respectively.

2. MATERIALS AND METHODS

To monitor the glacial velocities, different type of satellite data can be used: microwave (L-/C-band) Synthetic Apeture Radar (SAR) data of the satellites ESA, ENVISAT, ALOS-1/2, Sentinel-1/2 etc and optical data of the LANDSAT-5/7/8 and other commercial images. However, after processing the C-Band data (having wavelegth 4-5 cm) of ENVISAT and Sentinel-1, we found that a lot of errors and the off-glacial areas show un-necessarily high speeds, making the observed glacial speeds less reliable.

Also in the Karakoram Range, observed glaciers are large in size and most of the glaciers start from the elevations higher than 5000 meters (which are often covered with clouds). So, it is hard to find both master and slave images that are cloud free and have a short-temporal baseline, where one apply the feature tracking with the confidence that the surface features have been preserved in the slave image during a given time. It has been found that due to larger wavelength (23.6 cm) L-band ALOS data has potential to retain the coherency and is relatively good to study the glacial velocities in the Karakoram Range.

However, the pixel-offset tracking by L-band SAR is known to often generate “azimuth streaks” (Gray and Mattar, 2000; Kobayashi and others, 2009). In our ALOS-1/2 data, I observed the streaking pattern in many pairs of ALOS-2 but in few of ALOS-1. This is presumably because for ALOS-2, the data acquisition time is mostly in the local daytime, when there is a higher chance of irregularities in the ionosphere due to the higher background total electron content (TEC). Moreover, the streaking is more likely to appear in data collected close to or in the polar regions (Gray and Mattar, 2000). Furthermore, the amplitude of glacier displacement is essentially proportional to the temporal period between master and slave date, whereas the amplitude of azimuth streaks arises as a snapshot at the time of imaging. The observed glacier velocities are on the order of 100 m a^{-1} or more, and as such, the "azimuth streaks" did not substantially affect the observed velocity, and are within the errors of the observed velocities. But these streaks are possibly the main reason that ALOS-2 has more error in velocity than ALOS-1.

Following assumptions where made while extracting the glacial surface velocities using ALOS-1/2 data.

i. The ice penetration depth of L-Band microwave is around 15 m (Rignot and others, 2001). Here we assume uniform velocity through the ice layer near the surface down to this penetration depth.

ii. Typically there are clear modulations in seasonal velocity and the velocities are also the maximum in the ablation zone of a glacier. So the areas, that have shown clear seasonal modulations, and have the highest velocities compare to the upstream part were assumed as ablation zone, while the areas lying on the top of the ablation zone were considered as accumulation zone.

The processing method closely follows that of previous studies (Strozzi and others, 2002; Yasuda and Furuya, 2013; Abe and Furuya, 2015). The pixel-offset (feature/speckle) tracking algorithms are based on maximizing the cross-correlation of intensity image patches. The Gamma software package was used to process the SAR dataset (Wegmüller and Werner, 1997). For the ALOS-1 dataset, I use the Fine Beam Single/Dual (FBS/FBD) mode images along the ascending path. Their incidence angle at the center of the image is 38.7°, varying only around 4° from the near range to far range. The FBD data were oversampled in the range direction. The temporal baseline for ALOS-1 varies from 46–184 days, while for ALOS-2 it varies from 28–196 days (Table 1).

Table 1. ALOS-1/2 data used in this research. Bold areas show the reference pairs for which various feature tracking parameters are tested (YYYYMMDD).

Sensor	**Path/Look**	**Frame**	**Master Image**	**Slave Image**	**Mode**	**Time Span (days)**
PALSAR-1	523	690-700-710	**20061206**	**20070121**	**FBS-FBS**	**46**
			20070121	20070423	FBS-FBS	92
			20070423	20070608	FBS-FBD	46
			20070608	20070724	FBD-FBD	46
			20070724	20080124	FBD-FBS	184
			20080124	20080310	FBS-FBS	46
			20080310	20080425	FBS-FBD	46
			20080425	20080610	FBD-FBD	46
			20080610	20081211	FBD-FBS	184
			20081211	20090126	FBS-FBS	46
			20090126	20090613	FBS-FBD	138
			20090613	20090729	FBD-FBD	46

			20090729	20090913	FBD-FBD	46
			20090913	20091214	FBD-FBS	92
			20091214	20100129	FBS-FBS	46
			20100129	20100616	FBS-FBD	138
			20100616	20100801	FBD-FBD	46
			20100801	20100916	FBD-FBD	46
			20100916	20101217	FBD-FBS	92
	524	690-700-710	**20061223**	**20070207**	**FBS-FBS**	**46**
			20070207	20070810	FBS-FBD	184
			20070810	20070925	FBD-FBD	46
			20070925	20071226	FBD-FBS	92
			20071226	20080210	FBS-FBS	46
			20080210	20080327	FBS-FBS	46
			20080327	20080512	FBS-FBD	46
			20080512	20080627	FBD-FBD	46
			20080627	20081228	FBD-FBS	184
			20081228	20090212	FBS-FBS	46
			20090212	20090630	FBS-FBD	138

			20090630	20090815	FBD-FBD	46
			20090815	20090930	FBD-FBD	46
			20090930	20091231	FBD-FBS	92
			20091231	20100215	FBS-FBS	46
			20100215	20100703	FBS-FBD	138
			20100703	20100818	FBD-FBD	46
			20100818	20110103	FBD-FBS	138
			20110103	20110218	FBS-FBS	46
PALSAR-2	161/RF2_5	690-700-710	20140926	20150213	SM3	140
			20150213	**20150313**	**SM3**	**28**
			20150313	20150925	SM3	196
			20150925	20151204	SM3	70
			20151204	20160715	SM3	224
			20160715	20161202	SM3	140
			20161202	20170210	SM3	70
			20170210	20170714	SM3	154
			20170714	20171117	SM3	126
			20171117	20180518	SM3	182

	161/RF2_6	690-700-710	20141010	20141219	SM3	70
			20141219	**20150227**	**SM3**	**70**
			20150227	20150731	SM3	154
			20150731	20151009	SM3	70
			20151009	20151218	SM3	70
			20151218	20160729	SM3	224
			20160729	20161007	SM3	70
			20161007	20161216	SM3	70
			20161216	20170224	SM3	70
			20170224	20171020	SM3	238
			20171020	20171201	SM3	42
			20171201	20180601	SM3	182

For the ALOS-2 data, I use Strip Mode (SM3) imagery along the ascending path. In ALOS-1 data, Baltoro Glacier was covered by path 524, and the remaining glaciers were studied by using the data along path 523. In ALOS-2 data, the downstream area (from the terminus to ~25 km) of Baltoro Glacier was covered by the right-looking fine-beam mode (RLF2_5, with an incidence angle of 31.4°), whereas the ~25–49 km point of Baltoro and all of the other glaciers were analyzed by using data along the RLF2_6 mode with an incidence angle of 36.3°. The spatial coverage of each satellite is shown in Fig. 1.

To establish the best processing parameters, I selected the image pairs by considering two major factors. First, as the possibility of surface-feature preservation is high in winter, I selected pairs in the winter span (December to March). Second, pairs were selected with the shortest possible temporal baseline (46 days for ALOS-1; 28 and 70 days for ALOS-2 data) (Table 1). While various possibilities of search patch size and sampling interval were tested, the velocity has been derived with a search patch of 128 × 128 pixels (range × azimuth), with a sampling interval of 12 × 36 pixels. I set 3.0 as the threshold of the signal-to-noise ratio. The patches below this level were treated as missing data. The separation between satellite orbit paths and the effect of foreshortening over rugged terrain produces a stereoscopic effect known as an artifact offset (Strozzi and others, 2002; Kobayashi and others, 2009), which was corrected during the pixel-offset tracking with the use of the elevation-dependent correction. For the DEM, I used the Advanced Space-borne Thermal Emission and Reflection Radiometer (ASTER) global digital elevation model (GDEM) Version 2 (https://gdex.cr.usgs.gov/gdex/).

The velocity was calculated by following the parallel flow assumption (Joughin and others, 1998). The ASTER-GDEM was used to calculate the local topographic gradient unit vector. The detailed surface velocity field of each glacier is given in the supplementary materials file. To examine the spatial and temporal changes in velocity, I first determined the flow line at each glacier and then averaged the velocity data over the 200 × 200 m^2 area with its center at each flow line. The velocity errors were estimated by measuring the offsets on stable ground (non-glaciated area) (Pritchard and others, 2005; Yasuda and Furuya, 2013; Abe and Furuya, 2015), and they are 12–17 m a^{-1} for ALOS-1 data and 20–30 m a^{-1} for ALOS-2 data. Furthermore, I have divided the Baltoro and Siachen Glaciers into four regions: three regions of the main glacial channel (upstream, central, and terminal areas) plus the major tributary (e.g., Godwin-Austen for the Baltoro, Teram Shehr for the Siachen). The lengths of the sub-regions vary for each glacier, and the start and end of a region was set where I found some notable changes in velocity. This sub-division has not been done for the other three glaciers as there were no such notable changes in velocity along their length.

3. RESULTS

3.1. Baltoro Glacier

Of the five glaciers, the Baltoro is the only glacier whose detailed velocity evolution has been extensively studied (Quincey and others, 2009). About 35-km from its terminus lies Concordia, where the Godwin Austen Glacier comes down from the north and joins the main channel of Baltoro Glacier. The maximum velocity of ~180 m a^{-1} occurs just downstream of Concordia and from here the velocity gradually decreases downstream until the terminus (Fig. 2a). To examine the details, I divided the entire glacier into four regions from upstream to downstream (including the Godwin-Austen), and show the velocity evolution of each region separately (Fig. 2b-2e).

These spatial velocity patterns are mostly consistent with those observed by Quincey and others (2009), although some changes occurred after 2008. The summer speed-up signals are clear every year (Fig. 2a), but are more distinct in the downstream region. For example, the Godwin-Austen Glacier shows no significant summer speed-up signal except for the 2008 summer in its upstream part (Fig. 2e). As Quincey and others (2009) found for 2005, I see an extra speed-up in 2008, especially in the 10–35 km region (Figs. 2a, 2c, 2d). However, Quincey and others (2009) found that the terminal region (0–16 km) did not show significant changes in speed over the period 1992–2008. Except for the lowermost ~5 km, I observe a clear summer speed-up that is extended from the central region (Fig. 2d) in the ALOS-2 data.

Quincey and others (2009) observed faster speeds in the winter of 2007/2008. Using ALOS1/2 data, I also detected faster winter speeds in the same year (Fig. 2c) as well as in 2014/2015 and 2016 (Figs. 2b, 2c). However, the speeds are lower in 2008/2009 and 2009/2010. I can observe similar velocity changes in the Godwin-Austen Glacier, indicating faster speeds in the 2007/2008 and 2014/2015 winter seasons (Fig. 2e). We have found significant velocity increase in the upstream part in the fall of 2017 (Figs. 2a,b).

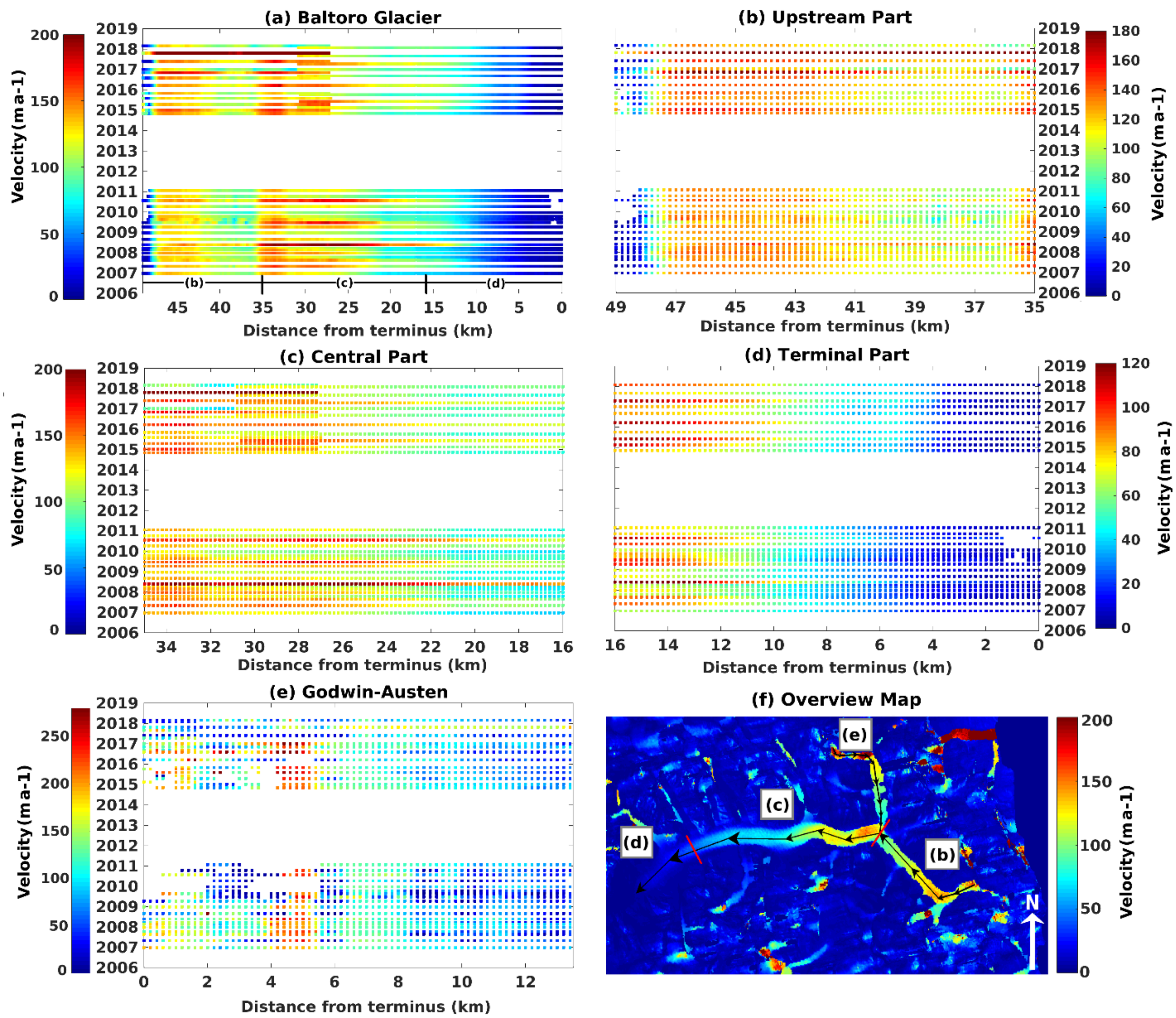


Figure 2. (a) Spatial and temporal surface velocity changes extracted along the entire length of the centerline of Baltoro Glacier. Expanded view of individual centerline segments of Baltoro Glacier; (b) upstream portion (49-35 km), (c) central portion (35-16 km), (d) terminal potion (16-0 km), and (e) the Godwin-Austen tributary (red lines indicate the boundary of each segmented centreline portion). The color scale in (a,b,c,d,e,f) indicates the scale for

the velocity of the glacier, its various parts and of the Overview Map. The numbers between the color scales and velocity figures indicate the time in years, and they are labeled at the start of the year, i.e., January. The detail of the pair dates are given in Table 1 and the velocity field of each pair is given in Figures S1. The green numbers indicate the master-slave images of beam having the Right Looking Fine beam mode (RLF2_6). (f) Surface velocity snapshot of Baltoro glacier derived from 23 December 2006 and 7 February 2007 data. The black line shows the profile line along which the velocity was calculated and arrows indicate the flow direction of the glacier.

3.2. Siachen Glacier

The velocity along the flowline of the Siachen Glacier decreases gradually from the upstream to the central region and reaches its maximum value between 30 and 40 km upstream from the terminus (Fig. 3a). In this region, I find clear seasonal and interannual variability. Below the 30-km point, the velocity steadily drops until the terminus of the glacier. The small velocity increases at 12 and 20 km probably arise from the tributaries. I again divide the entire glacier into four regions, and show the velocity evolution with a different color scale for each part (Fig.3b-e).

After 2014, the summer speeds over both the central and upstream regions appear slower than in previous years (Figs. 3a, 3c). For example, the 2015 summer speed in the central part is below 200 m a^{-1} and the upstream winter speed is slower than previous year's by more than 35%. The presence of seasonality in the upstream region is not clear because it often undergoes a decorrelation problem. This problem occurs because the summer snowfall at higher elevations can significantly change the surface features, creating difficulty for the offset-tracking algorithm, and thus making it difficult to measure the velocity (Figs. 3a,b). Nevertheless, in the 2009/2010 winter season, the results show a significant speed-up upstream, being nearly twice as high as all other years (Fig. 3b), and also show a speed-up over the central part (Fig. 3c). In contrast, the terminal region has its fastest speed in the 2015 summer, being 30% faster than the 2010 summer when the maximum speed occurs in the central part.

In contrast, Siachen Glacier's largest tributary, the Teram Shehr Glacier, shows very little interannual speed variations over the last 10 years (Fig. 3e). This nearly steady velocity distribution demonstrates that the observed velocity changes in the main trunk are not an artifact of the change of satellite from ALOS-1 to ALOS-2.

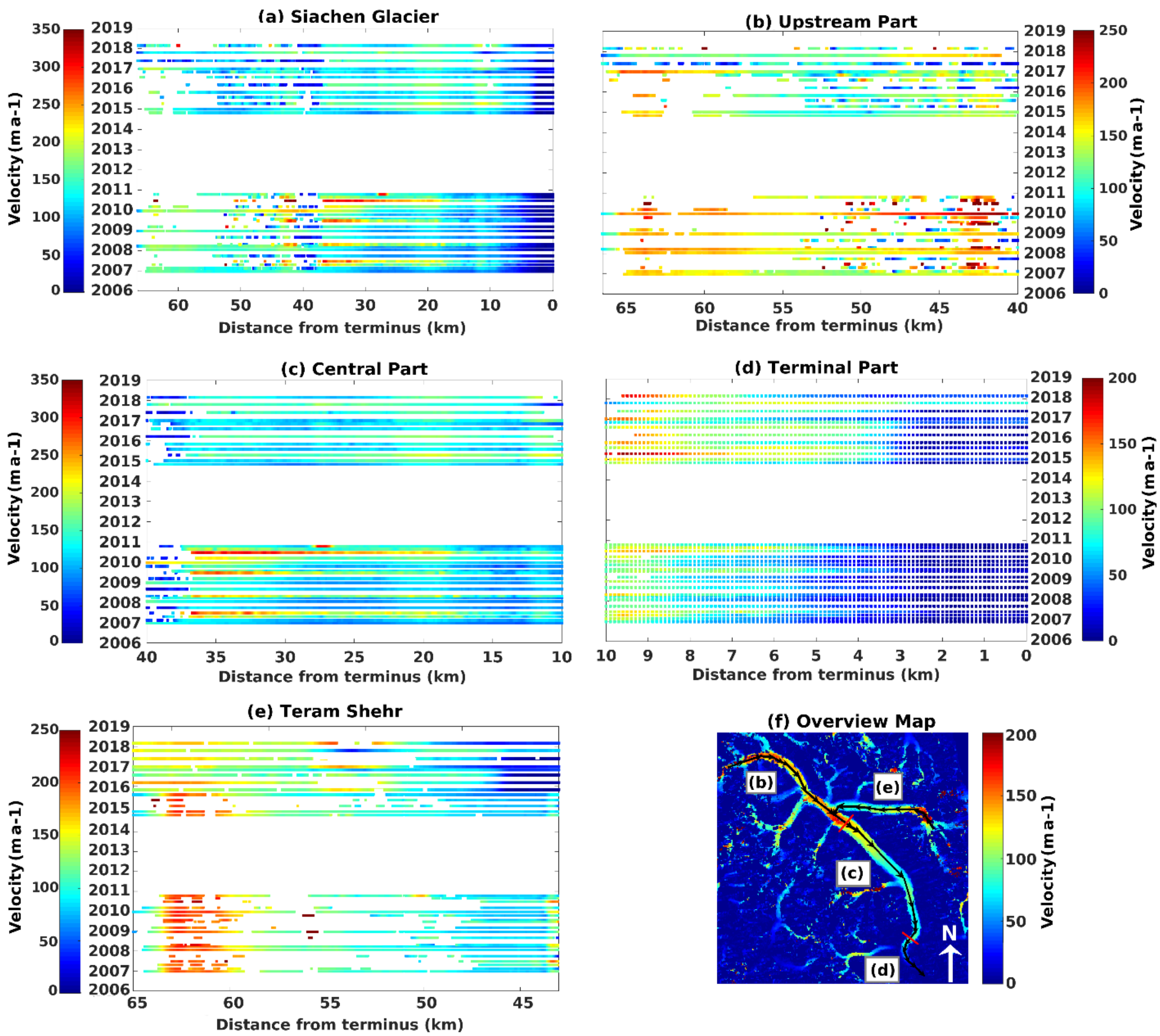


Figure 3. (**a**) Spatial and temporal surface velocity changes extracted along the entire length of the centerline of Siachen glacier. Expanded view of individual centerline segments of Siachen glacier; **(b)** upstream portion (66-40 km), **(c)** central portion (40-10 km), **(d)** terminal potion (10-0 km), and **(e)** the Teram Shehr Tribuatary (red lines indicate the boundary of each segmented centreline portion).The color scale in (a,b,c,d,e,f) indicates the scale for the velocity of the glacier, its various parts and of the Overview Map. The numbers between the color scales and velocity figures indicate the time in years, and they are labeled at the start of the year, i.e., January. The details of the pair dates are given in Table 1 and the velocity field of each pair is given in Figures S2. (**f**) Surface velocity snapshot of Siachen glacier derived from 6 December 2006 and 21 January 2007 data. The black line shows the profile line along which the velocity was calculated and arrows indicate the flow direction of the glacier.

3.3.Kundos Glacier

The Kundos Glacier has two tributaries. In the western tributary, the speed gradually decreases from the upstream region until the 23-km location (Fig. 4a), which is at the confluence of upstream tributaries. Then, between 23 and 18 km the speed increases, and downstream from this it gradually decreases. Upon passing another confluence around 14 km, it flows faster until about 9 km. After this zone, the speed decreases to the terminus. Thus, there are three segments of velocity pattern that are mostly controlled by incoming flow from tributaries (c.f. Bhambri and others, 2017; Jiskoot and others, 2017).

The velocity distributions of the eastern tributary are also segmented (Fig. 4b), and controlled by the presence of tributaries. The velocity has a maximum near the 30-km point, which is just below a confluence, then slowly decreases southward. A final increase occurs near 9 km at the main confluence of eastern and western tributaries (Fig. 4b). Below this point, most of the ice flow comes from the western tributary, with the western tributary preventing the inflow from the eastern tributary. Thus, the speed decreases to about zero at 9 km, and then abruptly jumps to the value for the western flow line. After this, both flow lines have the same speed. Concerning seasonal changes, both tributaries clearly show higher velocities in summer, although the amplitude of peak summer velocity is highly variable. For example, in the western tributary, the summer speeds in 2008 and 2015 are < 120 m a^{-1}, whereas those in other years are > 150 m a^{-1} (Fig. 4a). In the eastern tributary (Fig. 4b), the maximum summer speed is 100 m a^{-1} in 2008, but ~60% faster in the summer of 2010 and over 80% faster in 2015 (even over the 15–25 km portion, where no summer speed-up has been observed previously).

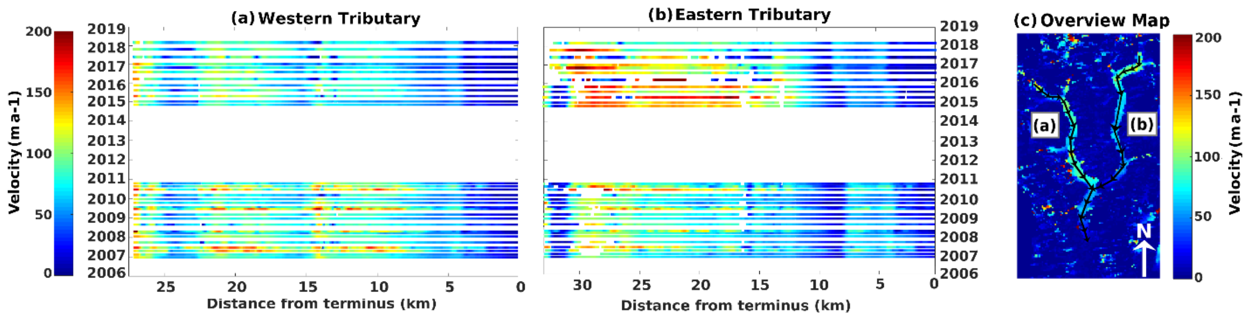


Figure 4. Spatial and temporal surface velocity changes extracted from the centerline of the a) Western Tributary and b) Eastern Tributary of Kundos Glacier. (**c**) Surface velocity snapshot of Kundos glacier derived from 6 December 2006 and 21 January 2007 data. The black line is a profile line along which the velocity was calculated and arrows indicate the flow direction of the glaciers. After the confluence, i.e., about the 0–9 km area of (a,b), the velocity was calculated along the same line for both parts. The color scale in (a,b) indicates the scale for the velocity of the glacier. The same scale has been used for the Overview Map as well. The details of the pair dates are given in Table 1 and the velocity field of each pair is given in Figures S3.

Since 2014, a higher velocity has been detected in the upstream part of the eastern tributary, regardless of the season**.** The 2014/2015 and 2015/2016 winter speeds are clearly faster by 30% or more than those in previous winter seasons.

3.4. Gasherbrum Glacier

Gasherbrum Glacier underwent a surge which started in fall 2005 and reached its peak velocity of 500 m a^{-1} during the summer of 2006 (Mayer and others, 2011; Quincey and others, 2011), and is now apparently in its quiescent phase (Fig. 5). Accordingly, the velocity steadily decreases from its upstream region to the terminus. Our data was collected since 2007, and complements the previous studies, showing the transition from the active phase to the quiescent phase in finer detail.

I observe a slow-down in the winter of 2006/2007, but then a significant acceleration in the summer of 2007, again reaching a velocity of 500 m a^{-1} (Fig. 5a). The latter peak speed was not reported in Quincey and others (2011), and indicates that peak velocities occurred both in the summers of 2006 and 2007. In the fall of 2007, the speed decreased to about 250 m a^{-1}, which is consistent with previous studies (Mayer and others 2011; Quincey and others 2011). After the active phase, the glacier becomes nearly stagnant in the lower ~10-km portion. Nonetheless, this glacier showed a summer speed-up in 2008 (Fig. 5a). However, the amplitude and spatial extent of the summer speed-up seem to have decreased since 2008.

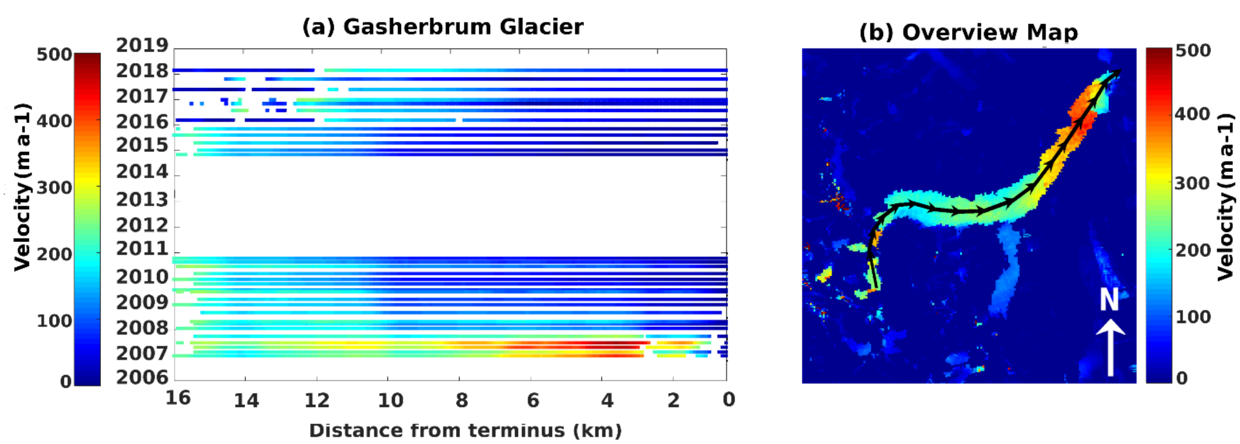


Figure 5. (**a**) Spatial and temporal surface velocity changes extracted from the centerline of Gasherbrum glacier. The color scale indicates the scale for the velocity of the glacier. The same scale has been used for the Overview

Map as well. The numbers between the color scale and velocity figure indicate the time in years, and they are labeled at the start of the year, i.e., January. (**b**) Surface velocity snapshot of Gasherbrum glacier derived from 6 December 2006 and 21 January 2007 data. The black line is the profile line along which the velocity was calculated and arrows indicate the glacier's flow direction. The details of the pair dates are provided in Table 1 and the corresponding velocity field of each pair is given in Figures S4.

3.5.Singkhu Glacier

Singkhu Glacier, at an elevation of 4500–5800 m a.s.l., is both relatively high and relatively short compared to the other glaciers. Due to its colder climate, the glacier may be less likely to have short-term velocity changes. However, our measurements show clear summer speed-up signals with inter-annual modulations, with the fastest speed occurring in the 2010 summer (Fig. 6a). In fact, there are two zones of faster summer velocity, one at 11–14 km and one at 3–10 km. As observed with Kundos Glacier, these two distinct velocity distributions are presumably due to confluences with a major tributary glacier.

As I found for the Siachen Glacier, I were not able to derive the upstream velocity in the summer season, which is probably due to the lower image correlation caused by snow accumulation in summer. In 2014–2015, however, I do not observe clear seasonal changes as observed previously, and the maximum summer speed declines, which agrees with that found for the Siachen and western tributary of the Kundos Glacier.

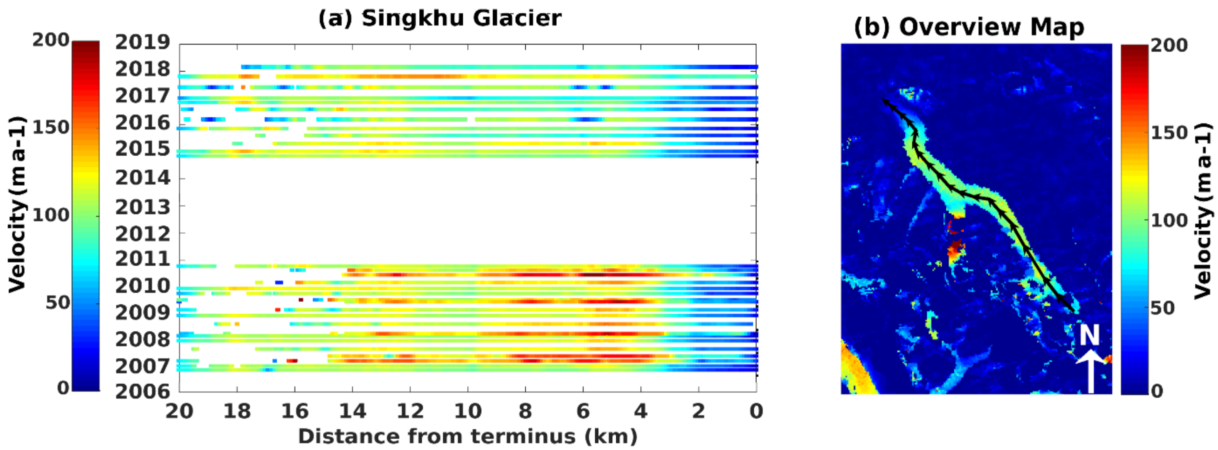


Figure 6. (**a**) Spatial and temporal surface velocity changes extracted from the centerline of Singkhu glacier. The color scale indicates the scale for the velocity of the glacier. The same scale has been used for the Overview Map as well. The numbers between the color scale and velocity figure indicate the time in years, and they are labeled at the start of the year, i.e., January. (**b**) Surface velocity snapshot of Singkhu glacier derived from 6 December 2006 and 21 January 2007 data. The black line is the profile line along which the velocity was calculated and arrows indicate the glacier's flow direction. The details of the pair dates are provided in Table 1 and the corresponding velocity field of each pair is given in Figures S5.

4. DISCUSSION

Research paper associated with this thesis published in “Journal of Glaciology” of Cambridge University Press

Title: Inter-annual modulation of seasonal glacial velocity variations in the Eastern Karakoram detected by ALOS-1/2 data

Authors: Muhammad Usman, Masato Furuya

Publication Year: 2018

Doi: 10.1017/jog.2018.39

4.1. Intra/interglacial spatio-temporal velocity variability at seasonal and inter-annual scales

Our derived velocity data show that both the amplitude and spatial extent of summer speed-up varies from year to glacier. For example, not all the glaciers have their maximum summer velocity in the same year. Moreover, maximum speeds within the same glaciers vary from year to year, which presumably indicates the variability of basal sliding, which is controlled by basal water pressure that depends not only on the input volume of surface meltwater, but also on the capacity of the subglacial hydraulic system at each individual glacier.

Concerning the larger summer speed-ups, Quincey and others (2009) found an anomalously fast summer speed-up in 2005 at Baltoro Glacier, which was interpreted to indicate additional basal sliding due to the larger volumes of meltwater made available by the deep snowpack in the preceding winter. I found two notable summer speed-ups: one in 2008 at Baltoro Glacier and one in 2010 in the central part of Siachen Glacier. Moreover, I found additional summer speed-up signals, such as on the terminal part of Siachen in 2015 and on the eastern tributary of Kundos in 2015.

All summer speed-ups may be due to a large-volume surface meltwater pulse each spring or summer (c.f. Müller and Iken, 1973; Iken and Bindschadler, 1986). However, this argument is implicitly based on two arguable assumptions: (1) the large surface meltwater pulse is limited to each local glacier at different times, and (2) surface meltwater can efficiently reach the bed and broadly increase basal water pressure during the corresponding same summer.

For the first assumption one should consider the possibility of heterogeneous climate over the studied area. For such analyses, as surface meteorological data is probably infeasible, one may need to run downscaling analysis of large-scale global numerical weather data (e.g., Bieniek and others, 2016), considering the complex topography and large elevation differences that can affect the surface weather conditions. However, for the Kundos Glacier, where the close proximity of its tributaries suggests similar surface meteorological conditions, I found that the extra summer speed-up occurred only in the eastern tributary in 2010 and 2015, whereas the western tributary has slowed down since 2015. Instead, the eastern tributary's 2015 speed-up may be a sign of surge-initiation, which I will discuss in more detail below. This case suggests that the external meteorological forcing at each glacier does not directly control the summer speed in the same year.

For the second assumption an efficient passage of meltwater may not be necessary. For example, the summer velocity increases in 2008 at Baltoro, 2010 at Siachen, and 2015 at the eastern tributary of Kundos Glacier were all preceded by relatively fast speeds in the previous winter in their upper regions (Figs. 2b, 2c, 3b, 3c, and 4b). These observations indicate that the extra basal sliding does not necessarily start in the summer, but instead already during the preceding winter in the upstream regions. Although the cause of this upstream speed-up is unclear, once the speed-up occurs it could help surface meltwater in the following summer to more broadly increase the basal water pressure over the entire glacier. For example, the winter increase in basal sliding would generate more space on the leeward side of bedrock bumps (Kamb, 1987; Schoof, 2010), making the basal water cavities more widely connected when large volumes of meltwater become available in spring to early summer (Iken and Truffer, 1997). Also a recent speed up singal observed in the winter of 2016 and in the fall of 2017 in upstream part of Baltoro Glacier

(Fig. 2b) would possibly follow a speed up in the summers of 2017. There, it is important to continue focus on this glacier.

The question that follows is how the upstream speed-up can be initiated in winter. The winter speed-up in the upstream that precedes a larger summer speed-up is only observed once every several years. The interim time period may be a “recharge time” for the upstream subglacial hydraulic system to generate higher basal water pressure. Unlike the downstream ablation zone, the upstream region generally does not have high volumes of surface meltwater input every summer, and thus it would take longer time to store basal water by either pressure-melting of ice like the Svalbard-type thermal regulation (e.g., Murray and others, 2003) or interstitial englacial water generated by strain heating (Aschwanden and Blatter, 2005) and their gravity-driven movement as suggested by Irvine-Fynn and others (2006). Although ice creep will generate locally isolated cavities with high-pressure water, the extra speed-up in winter requires that those cavities become broadly connected, which would also take over one year. In this way, the “recharge time” for the upstream winter speed-up would be several years, ultimately leading to the larger summer speeds every several years. The extra summer speed-up preceded by upstream winter speed-up might be better viewed as “surge-like” events reported by Bhambri and others (2017). However, a number of uncertainties remain in the upstream glacial hydrology (Irvine-Fynn and others, 2011).

The recent decrease in summer flow speed at Siachen, the western tributary of Kundos, and the Singkhu Glaciers may have occurred because the basal hydraulic system has become more efficient, perhaps due to a longterm larger influx of meltwater reducing the basal water pressure. Also, given the same amount of surface melt, there is a possibility that the efficiency of hydraulic system may change from time to time and may cause variable basal water pressure on the seasonal inter-annual scales that ultimately result in the diverse patterns of surface velocity. At the Godwin-Austen Glacier, the absence of a summer speed-up may indicate that either no meltwater could reach the base of glacier (e.g., due to colder ice in the upstream region) or that the morphology of the underlying bed is such that the summer melt does not have an impact on the ice motion.

4.2. Role of hydrological and thermal control mechanisms for flow instability in the Karakoram Range

Surge-type glaciers dominate the Karakoram Range (e.g., Hewitt, 1969, 2007, 2011), but their generation mechanism remains uncertain (e.g., Harrison and Post, 2003; Jiskoot, 2011; Sevestre and Benn, 2015).

To explain differences in surge behaviour, Murray and others (2003) proposed the Alaskan-type hydrological regulation and Svalbard-type thermal-regulation mechanisms for the generation of high-pressure basal water that could drive an active surge. Considering the dynamics and seasonality, for thermally-controlled surges, the initiation and termination phase lasts for several years before and after the peak phase, respectively (Quincey and others, 2015); and they are independent of the seasons (Jiskoot, 2011). On the other hand, hydrologically-driven surges are characterized by a rapid acceleration and deceleration (Quincey and others, 2015) and have tendency to initiate in winters (Raymond, 1987; Harrison and Post, 2003) and terminate in summers (Björnsson, 1998). The two models are also entirely different in terms of the origin of basal water. In the hydrological model, the water mainly originates from surface melt, while in the thermal-regulation model it is generated by pressure melting and frictional heating by the accelerated basal motion without input of surface meltwater. Recent observations, however, suggest that any of these mechanisms cannot be generalized to a study area containing multiple surge-type glaciers (Jiskoot and Juhlin, 2009; Quincey and others, 2015).

For the Karakoram, Quincey and others (2015) proposed that the surges are part of spectrum from normal slow flow to permanent fast flow, governed by hydrological and basal thermal processes. Moreover, Paul (2015) and Bhambri and others (2017) argued that Karakoram glacier surges are only marginally affected by the external climate forcing. We find similar behavior in our study area. In particular, Siachen and the eastern tributary of Kundos have shown surge/mini-surge initiation signals independently from each other, indicating that these events are probably not controlled by climate forcing, but rather are the result of internal mechanisms of the glaciers (Jiskoot, 2011; Raymond, 1987).

Of the five glaciers examined in this study, only Gasherbrum Glacier was surging, and in transition to its quiescent phase. In combination with previous studies that covered this glacier's earlier phases (Quincey and others, 2011; Mayer and others, 2011), we find that the surface velocities during the active phase turn out to be seasonally modulated with their peak flow in summer. This observation may have an important implication for the Karakoram surge mechanism, particularly because similar seasonal modulation during the active phase has also been found at two glaciers in West Kunlun Shan, northwestern Tibet (Yasuda and Furuya, 2015) and at Hispar Glacier in central Karakoram (Paul and others, 2017). In particular, I consider that the seasonal modulation of the peak velocity amplitude is possible evidence for the role of surface meltwater input to maintain the active phase of polythermal surge-type glaciers; polythermal structure has been assumed by Quincey and others (2015), too. As observed in

Karakoram glaciers by Bhambri and others (2017), an active surge could generate surface crevasse probably due to the increased longitudinal extensional stress over the glacier through which surface meltwater could easily reach the glacier bed and thereby further enhance the basal water pressure. This process would then help maintain the seasonally modulated active phase. Although it was not a seasonal modulation, Sund and others (2014) suggested a similar process for the surge development at the polythermal surge event of Nathorstbreen glacier system, Svalbard.

For Siachen Glacier, the faster flow region initiated in the upstream part of the glacier in early 2010 may have propagated downstream. Although it is not a surge, this is reminiscent of a "mini-surge" or "glacier pulse" observed in Alaskan glaciers (Kamb and others, 1985) and Karakoram glaciers (Bhambri and others, 2017), or an incomplete surge development as observed for Svalbard glaciers (Sund and others, 2009). Because no other examined glaciers revealed similar acceleration in 2010, it is probably not due to any anomalous external climate forcing in 2010, but rather due to local and internal mechanisms at the glacier. For the onset part of Siachen glacier (Figure 3b), before the initiation of mini-surge in the winters of 2010-2011, we can observe a gradual increase in velocity for the two consecutive winter seasons (i.e. 2007-2008 and 2008-2009), then there is a slight retardation in the winters of 2009-2010. In Karakorum Range, when surface velocity is taken into consideration, thermally-controlled surge develops gradually over the years (Quincey and others, 2011) and it can initiate at any time (Jiskoot, 2011). While the hydrologically-controlled surge has tendency to initiate in winter (Raymond 1987; Harrison and Past, 2003) and tends to terminate in summer (Björnsson, 1998). Considering the kinematic and seasonal characteristics of both surge types, we cannot conclude with confidence about the possible surge mechanism for the upstream part of Siachen glacier. However, on the basis of available data, we can observe these two factors: the winter initiation and significant velocity reduction in ALOS-2 data. These factors give indication of hydrological-controlled surge mechanism. Also, for the terminal part of Siachen and upstream part of Eastern tributary of Kundos glacier in ALOS-2 data, we can observe an acceleration in the winter before the maximum speed event in the summers of 2015 and then there is a velocity decline in the fall of 2015 and then there is an acceleration in the winter of 2015. This seems analogues to the surge velocity behavior of Hispar Glacier in Karakorum Karakorum Range, where Paul and others (2017) also notice a rapid decline in velocity immediately after the maximum speed event in the summer of 2015, also, the velocities again increased significantly in winters of 2015-2016 (i.e. after the sudden velocity decline event). Based on the available data, we may regard this event as Alaskan-type glacial surge.

The eastern tributary of Kundos Glacier may have started to surge recently (Fig. 4b). Acceleration occurred in the winter before the maximum speed in the summer of 2015, and this was followed by a velocity decline in the fall of 2015, and another acceleration in the winter of 2015. Even though it may not be a surging episode, the behavior is similar to the surge velocity behavior of Hispar Glacier in the Karakoram, where Paul and others (2017) also noticed a rapid decline in velocity immediately after the maximum speed event in the summer of 2015, with the velocities again increasing significantly in winter 2015–2016 (i.e., after the sudden velocity decline). As Paul and others (2017) suggested for the Hispar Glacier surge, the recent speed-up data at the eastern tributary of Kundos Glacier seems to conform to the Alaskan-type glacial surge model.

Particularly for Alaskan glacier surges, it is known that active phase often initiates in winter (Raymond, 1987). Although there are few reliable reports in Karakoram, Bhambri and others (2017) also mentioned a couple of active surges initiated in winter. Furthermore, all the mini-surge events in our study area start in winter as I suggested in the previous section. As winter should have a less efficient basal-drainage system that can more easily generate higher basal water pressure, the initiation in winter may involve subglacial water pooling over a multi-year period due to a combination of basal pressure melting, geothermal heat flux at the bed, and/or englacial water storage that can persist over winter (Lingle and Fatland, 2003; Abe and Furuya, 2015).

Although we have separated additional summer speed-up from apparent surging events in this paper, they should lie within the continuous flow-instability spectrum, which much broader than the conventional simple classifications such as “surge”, “pulse”, “mini-surge”, and decadal “Svalbard-type” (Jiskoot, 2011; Herreid and Truffer, 2016). Our essential argument to include both behaviours in this spectrum is that both events are likely controlled by subglacial and possibly englacial hydrology in the upstream region.

4.3. Implications for the Karakoram Anomaly

Copland and others (2011) suggested that the recent positive mass balance in the Karakoram played a role in the doubling in number of new surging episodes during the 14-year-period before and after 1990. However, the number of surge-type glaciers in their active phase has decreased since 1999, which is hard to explain only by the changes in mass balance, thus suggesting complexity in the surging mechanisms (Rankl and others, 2014). Out of five glaciers in our study area, we have studied high surge/minisurge events in three glaciers (Gasherbrum, Siachen and Eastern Tributary of Kundos Glaciers) which indicates that considering the number of surge events, Karakoram Range is more dynamic than assumed by Rankl and others (2014).

Also, in the case of both eastern-western tributary of Kundos Glacier, which lay in a relatively close proximity, suggests that (the western is decelerating and eastern is surging in ALOS-2 data) the velocities are either controlled by their local and internal mechanisms or the climate is so heterogeneous that each individual glacier has unique climate conditions. If the later is true than we need new approaches to study the Karakoram Glaciers as the diverse dynamic of the these glaciers complicates efforts to establish the link between glacial dynamics and the climate forcing (Bhambri and others, 2017).

We have detected Alaskan-type characteristics for the surge/minisurge events in our study area. As the Jiskoot (2011) proposed that the Alaskan-type surge results, where the basal-hydraulic system may change from conduit dominated to distributed link cavity system. The cause of this switch was proposed that the collapse of conduit. One can make two assumptions for the conduit collapse: i) the conduit my outgrow itself (Jiskoot, 2011) ii) conduit may collapse due to the over-burden pressure.

For the first assumption, there can be two possibilities: either the uniform basal water flow over the time may result in significant expansion of the conduit that it cannot support the over-burden pressure and may collapse or there is sudden increase in glacial outflow that can result in significant expansion of the conduit that it cannot support the over-burden pressure and may collapse. For the former, uniform outflow gives indications that the mass of the glacier will not change with the passage of time. For the later however, one may conclude that the significant glacial melt (i.e. mass loss) can produce more water that can flow through basal conduit and outgrows it significantly and results in it collapse. The the accelration related to the ice mass loss was observed in the Greenland ice sheets, where Zwally et al. (2002) concluded that the glacial sliding is enhanced by rapid migration of surface meltwater to the ice-bedrock interface.

For the second assumption, one may attribute the significant mass increase over the time to the overburden pressure increase. This point seems more close to the Karakoram Anomaly. However, as now we have concluded that Alaskan-type surge mechanism dominates the Karakoram Range, so the observed speed up can be attributed to either mass stability, mass loss or even mass gain. It is important to note that the local people claim that there is significant mass loss in Karakoram Glaciers. As they observe that glacial terminus are now retreating, also they say there is a significant increase in glacial outflow that is resulting in the formation of new lakes. On the basis of these results and observations, it is very important to conduct the field observations and use some new approaches to examine that is Karakoram "Anomaly" is actually an anomaly.

4.4. **Limitations and recommendations**

Our velocity observations demonstrate that the studies Karakoram glaciers are much more dynamically variable in terms of both spatial and temporal evolution than previously reported. Indeed, as our velocity maps are derived from satellite imagery and thus give an average between the acquisition times, the actual velocity changes could be even more dynamic (c.f. Armstrong and others, 2017). However, our study is confined to surface velocity data only, and we have not considered independent data such as local weather, surface meltwater volume, basal water pressure, and basal topography. Thus, the mechanisms we argued here are speculative. Nonetheless, I believe that the observed diverse range of surface velocity variations will have important implications for future studies, not only at glaciers in Karakoram but also other regions in the world.

To prove or reject our hypotheses I need not only to extend the time coverage of velocity observations, but also to collect local weather data and analyze the spatio-temporal evolution of weather data. Further testing should also involve direct measurements of basal water pressure in the upstream region over several years. Besides the collection of ground-based meteorological and hydrological observations and their comparative analysis with glacier dynamics at selected sites, modeling studies that can couple glacial hydrology with ice dynamics would also be useful. While there are already some pioneering studies that could quantitatively explain the observations at both mountain glaciers and ice sheets (e.g., Kessler and Anderson, 2004; Pimentel and Flowers, 2010; Pimentel and others, 2010; Hewitt, 2013; Kyrke-Smith and others, 2014; Hoffman and Price, 2014; Hoffman and others, 2016), most studies have focused on the summer speed-up signal attributable to the meltwater input in the same year. Any future model that couples subglacial hydrology with ice dynamics may need to include not only the downstream ablation zone, but also the upstream accumulation zone, and to extend the time coverage for development of the hydrological system from years to decades.

5. CONCLUSIONS

We applied offset tracking techniques to most of the available L-Band SAR data of ALOS-1/2 to determine the surface-velocity trends of five glaciers in the Eastern Karakoram Range over the period 2007–2015. The glaciers show clear seasonal changes in velocity, but with interannually modulated amplitudes and varied spatial-extents. Trends for each glacier appeared independent of the others, indicating that their velocities are probably controlled by local and internal mechanisms. The input from tributaries created upstream bands of lower speed, leading to a segmentation of the velocity trends along the length of the glaciers.

We found that Baltoro Glacier had a velocity increase in the summer of 2008 in its central part and an overall speed-up in its upstream part in 2015–2016, whereas Siachen, Singkhu, and the eastern tributary of Kundos Glaciers had a speed-up in the summer of 2010. For Siachen Glacier, the longest tributary, Teram Shehr, had a nearly steady velocity distribution throughout the study period. Since 2014, the western tributary of Kundos Glacier slowed down, while in the eastern tributary speed-up signals were observed over the upstream part, starting in winter and reaching a peak velocity in the summer of 2015. For Siachen Glacier, the upstream and central parts slowed, while the terminal part had an overall acceleration. On the basis of available data, the velocity changes observed at Siachen and eastern tributary of Kundos Glaciers are interpreted as a mini-surge, suggesting the Alaskan-type surge mechanism. The peak velocities of the Gasherbrum surge were seasonally modulated with two maxima in the summers of 2006 and 2007.

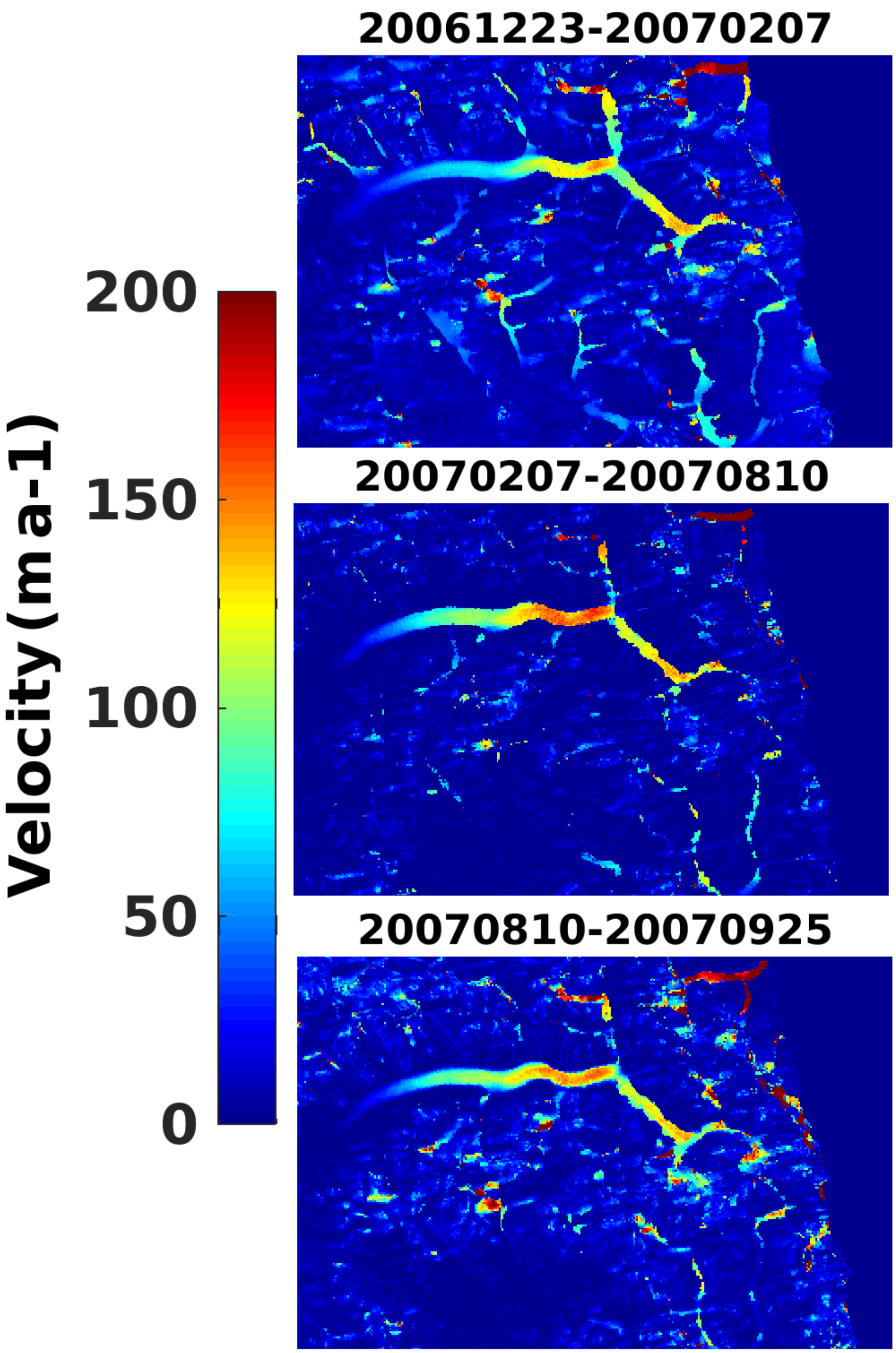
Out of five glaciers in our study area, we have studied high surge/minisurge events in three glaciers (Gasherbrum, Siachen and Eastern Tributary of Kundos Glaciers) which indicate that considering the number of surge events, Karakoram Range is more dynamic than assumed by

Rankl and others (2014).Also, in the case of both eastern-western tributary of Kundos Glacier, which lay in a relatively close poximity, suggests that (the western is decelrating and eastern is surging in ALOS-2 data) the velocities are either controlled by their local and internal mechanisms.

Concerning the mechanisms for flow variability, we argue that the intra/inter spatial and temporal changes in englacial/subglacial system of each glacier may vary the input of surface meltwater to the base and result in diverse velocity patterns in these glaciers. Our surface velocity findings show intriguing variations, but their temporal resolution and duration are not sufficient to understand the processes of glacier dynamics fully. To more quantitatively interpret the detailed interannual modulation of seasonal velocity variations, one needs to acquire regional high-resolution meteorological and hydrological data as well as run numerical modeling that can fully couple ice dynamics with multi-year englacial and subglacial hydrology.

6. ADDITIONAL FIGURES

Surface velocity field of Baltoro Glacier from 2007 to 2016 (Figures S1)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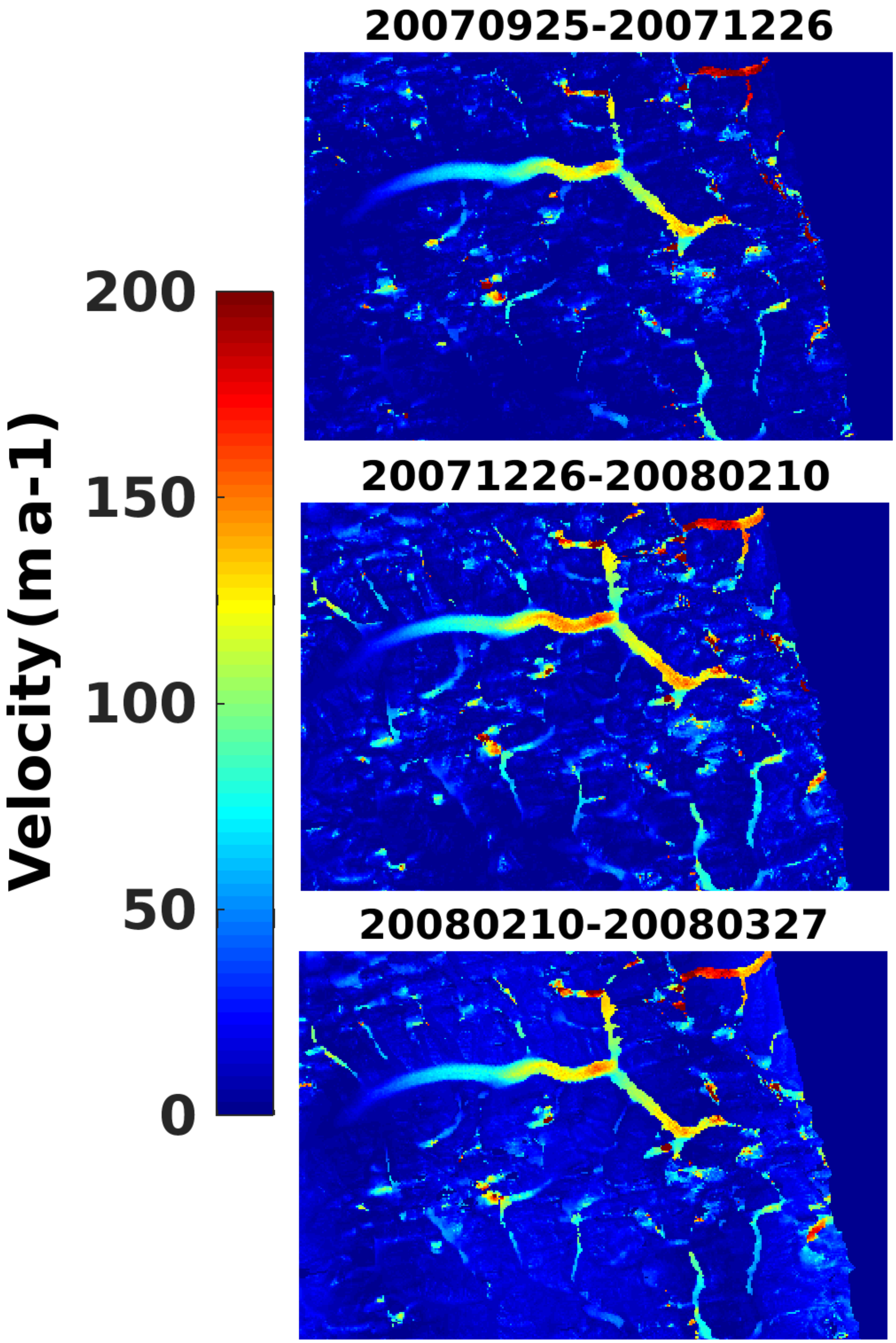
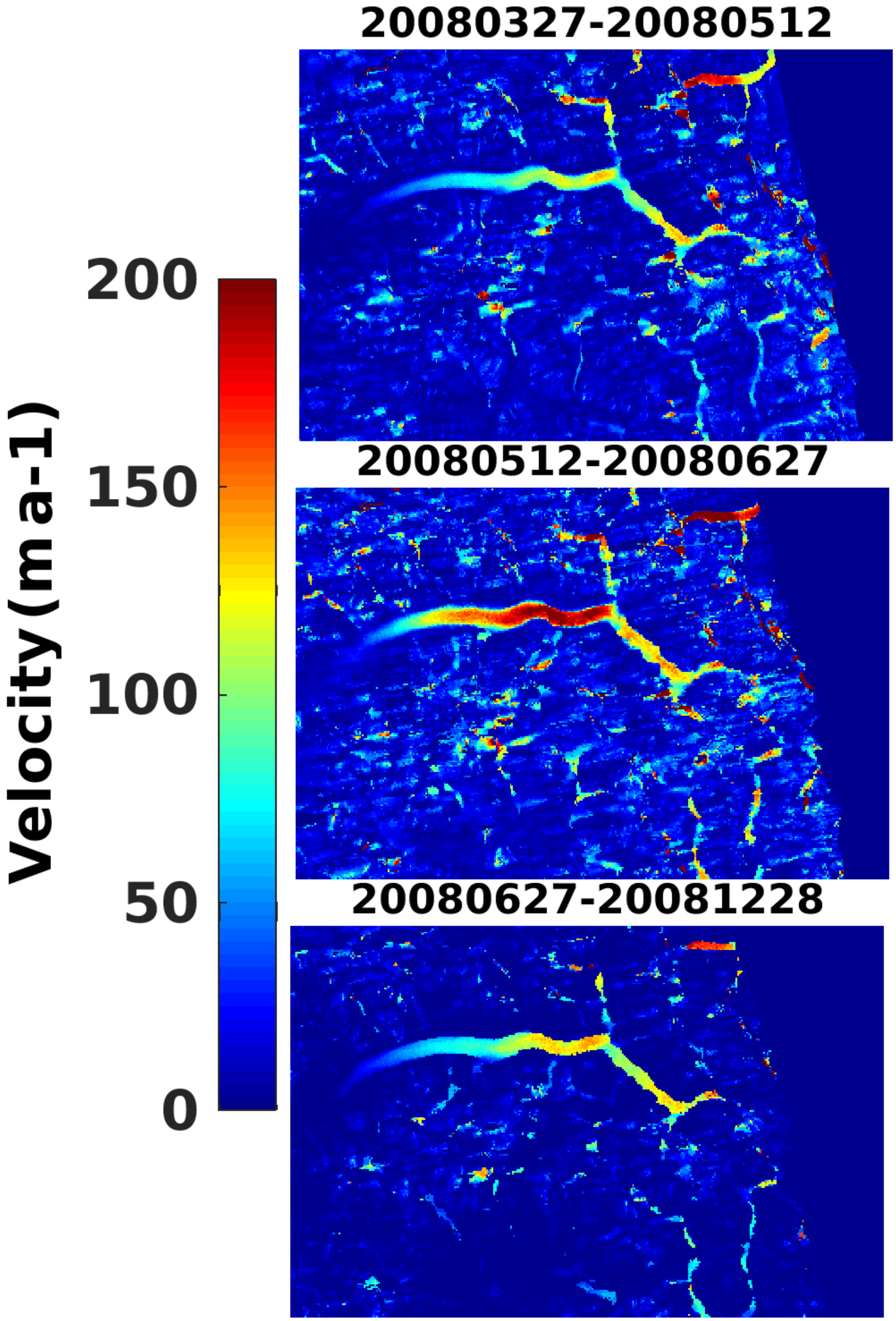
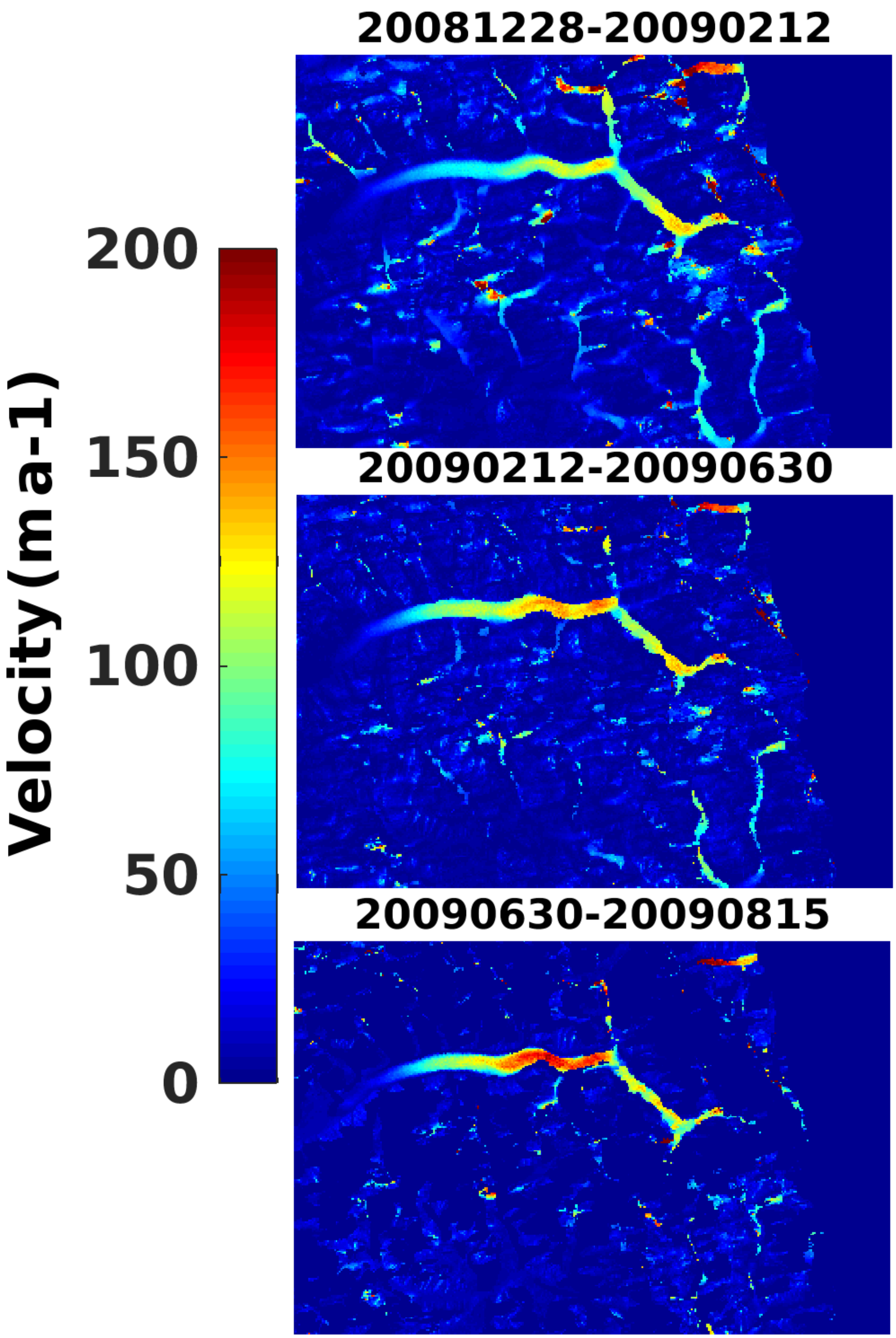
20070207-20070810
20070810-20070925
200
150
100
50
0
Velocity (m a-1)



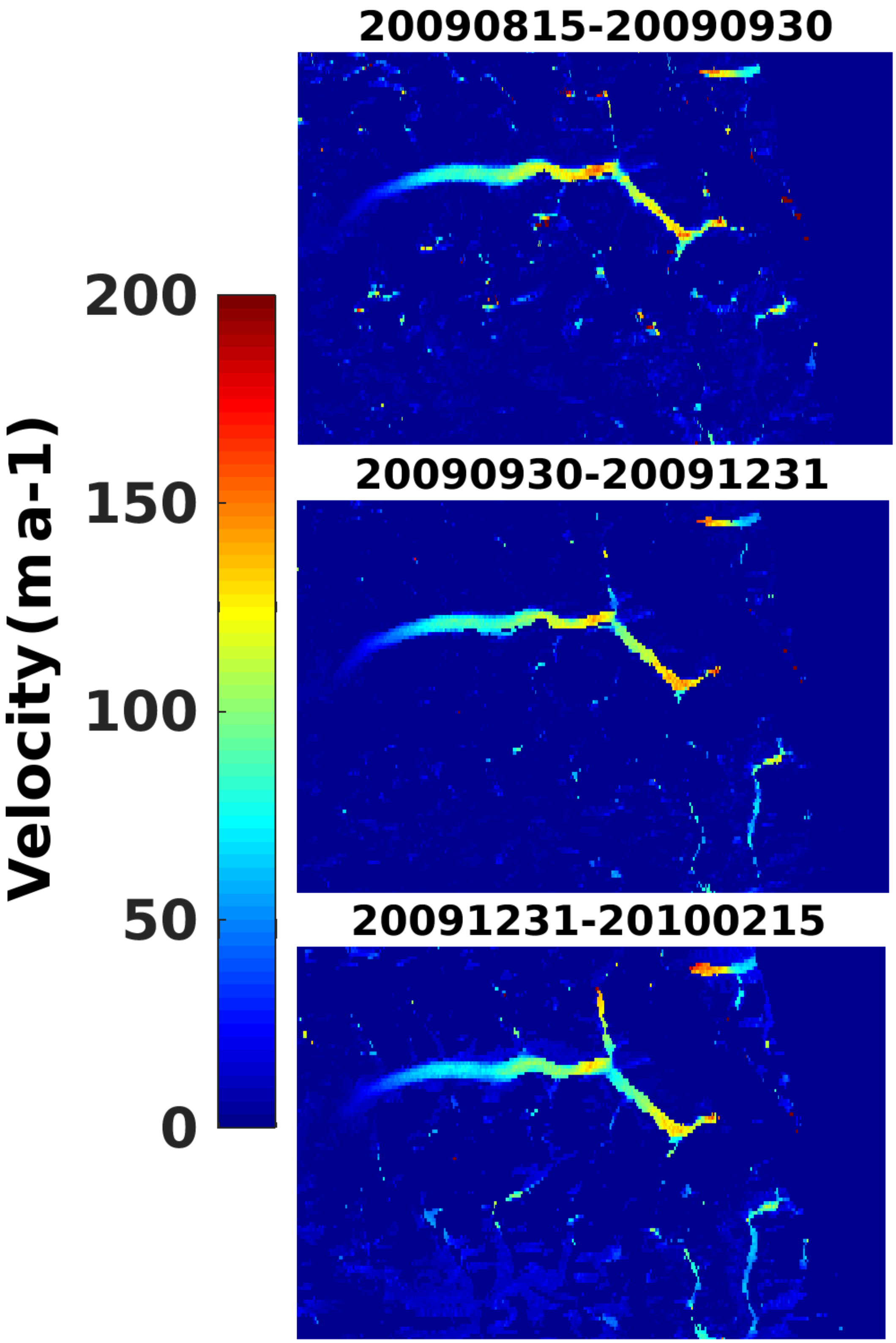
20070925-20071226
20071226-20080210
20080210-20080327
Velocity (m a-1)
200
150
100
50
0

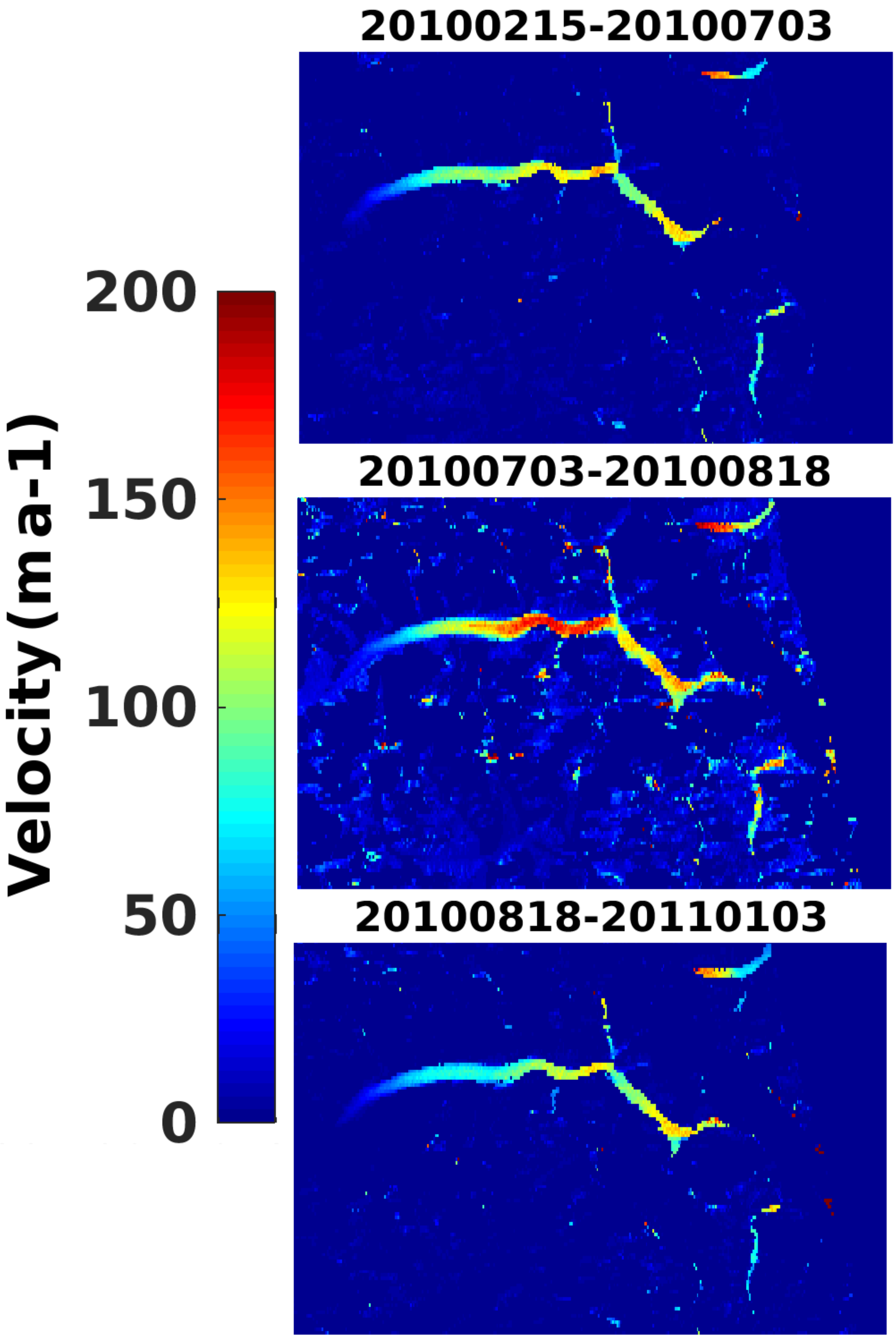


Velocity(m a-1)
200
150
100
50
0
20080327-20080512
20080512-20080627
20080627-20081228

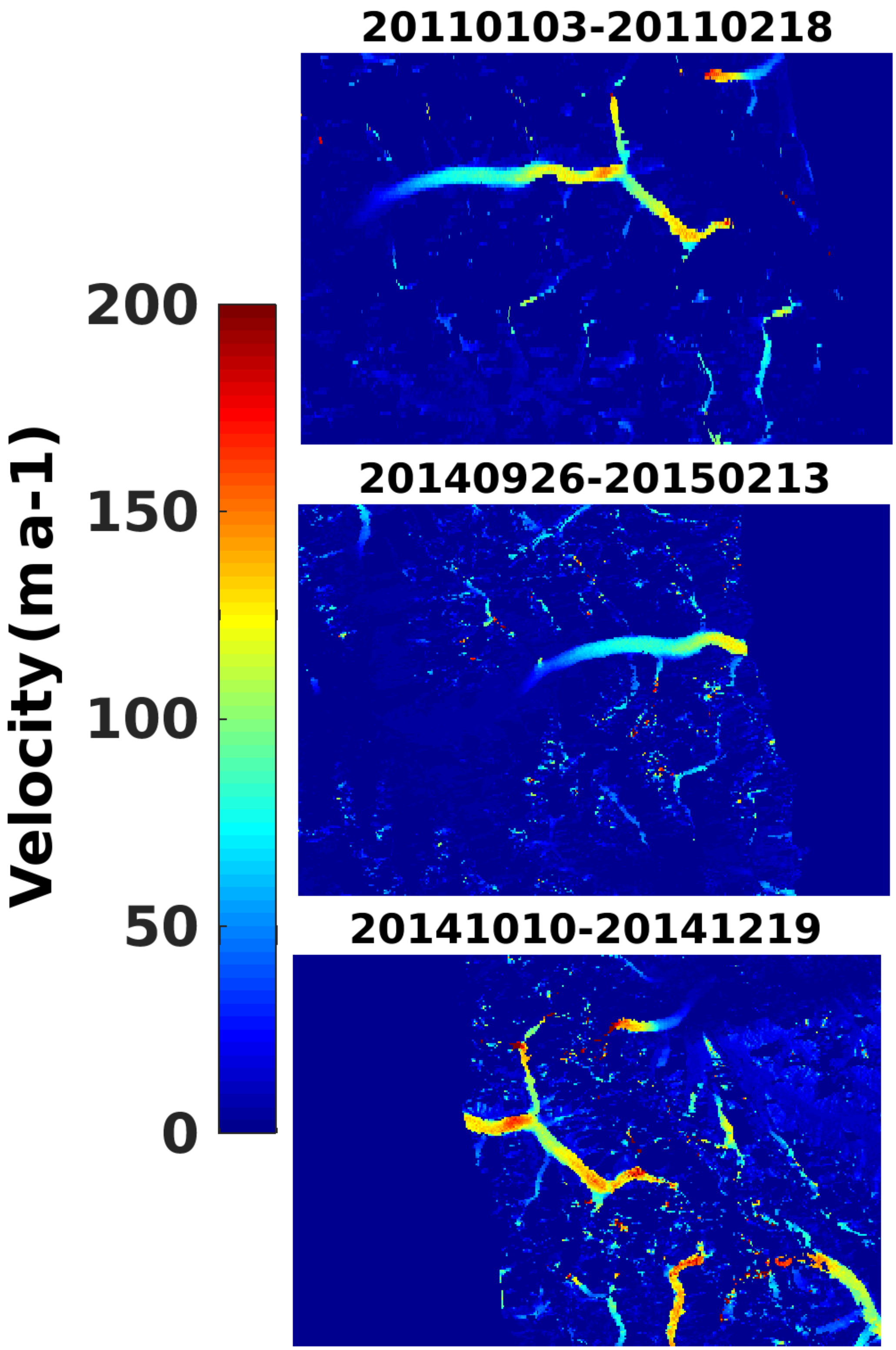


20081228-20090212
20090212-20090630
20090630-20090815
Velocity (m a-1)
200
150
100
50
0

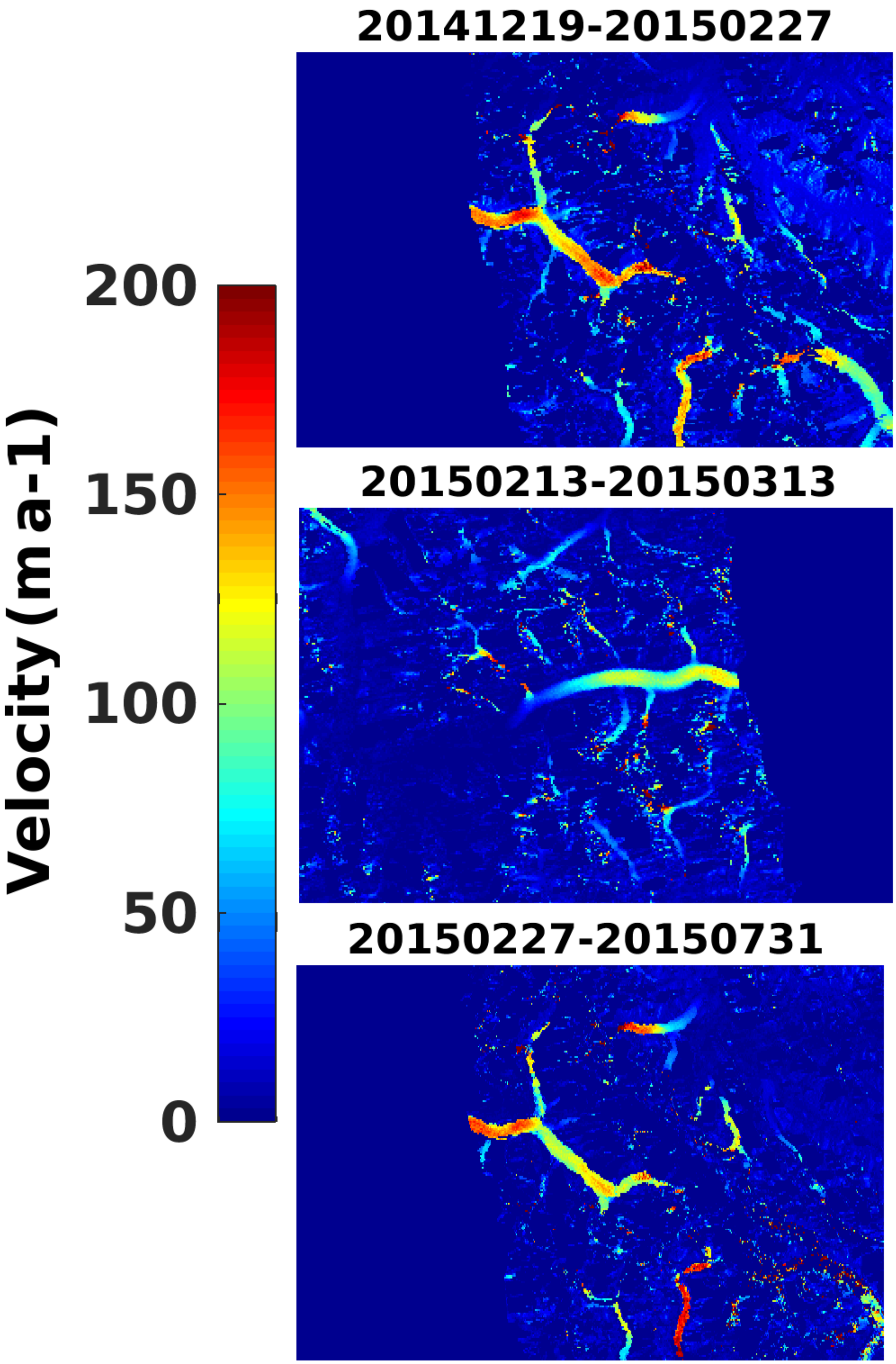

20090815-20090930
20090930-20091231
20091231-20100215
Velocity (m a-1)
200
150
100
50
0

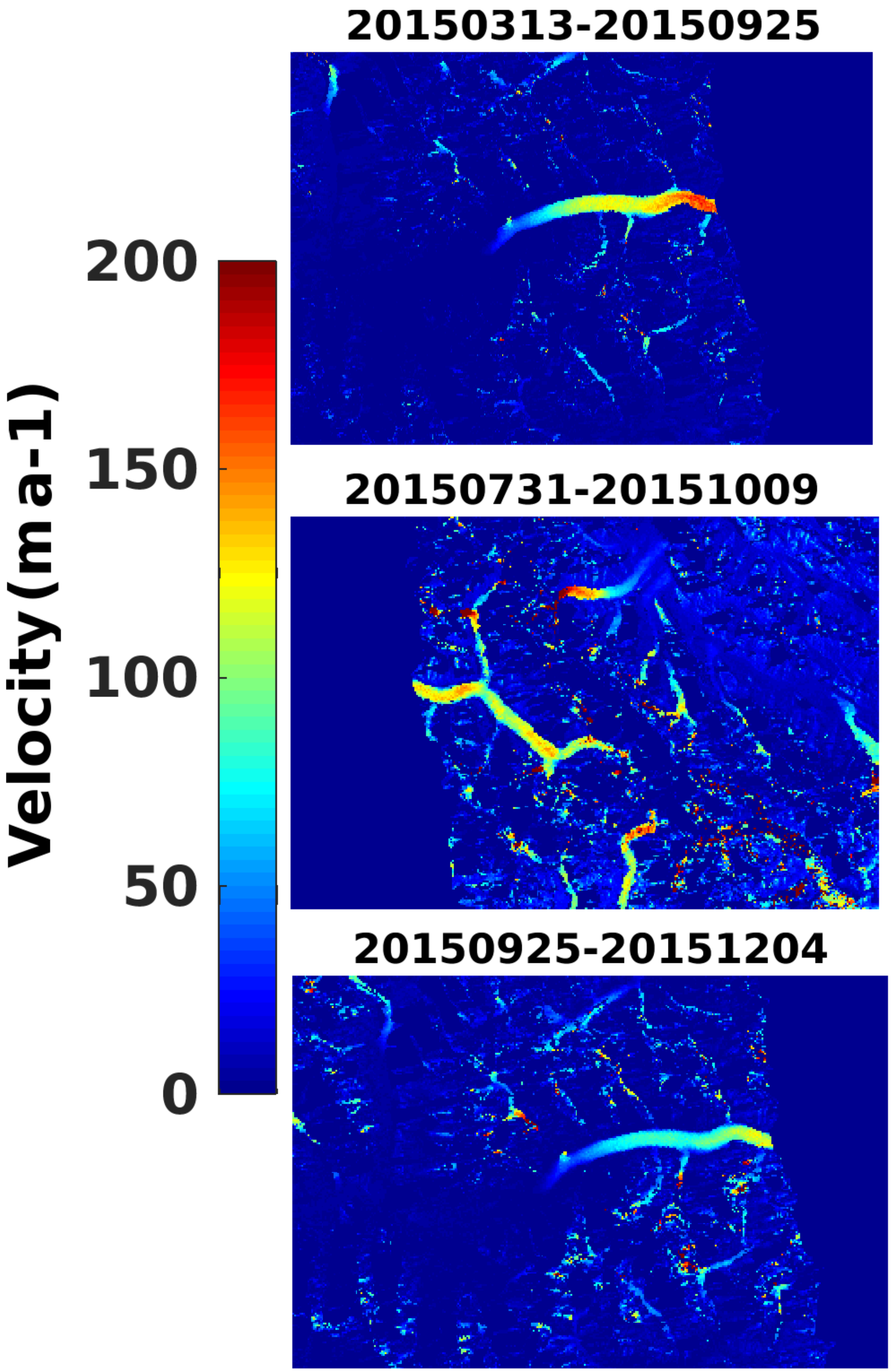


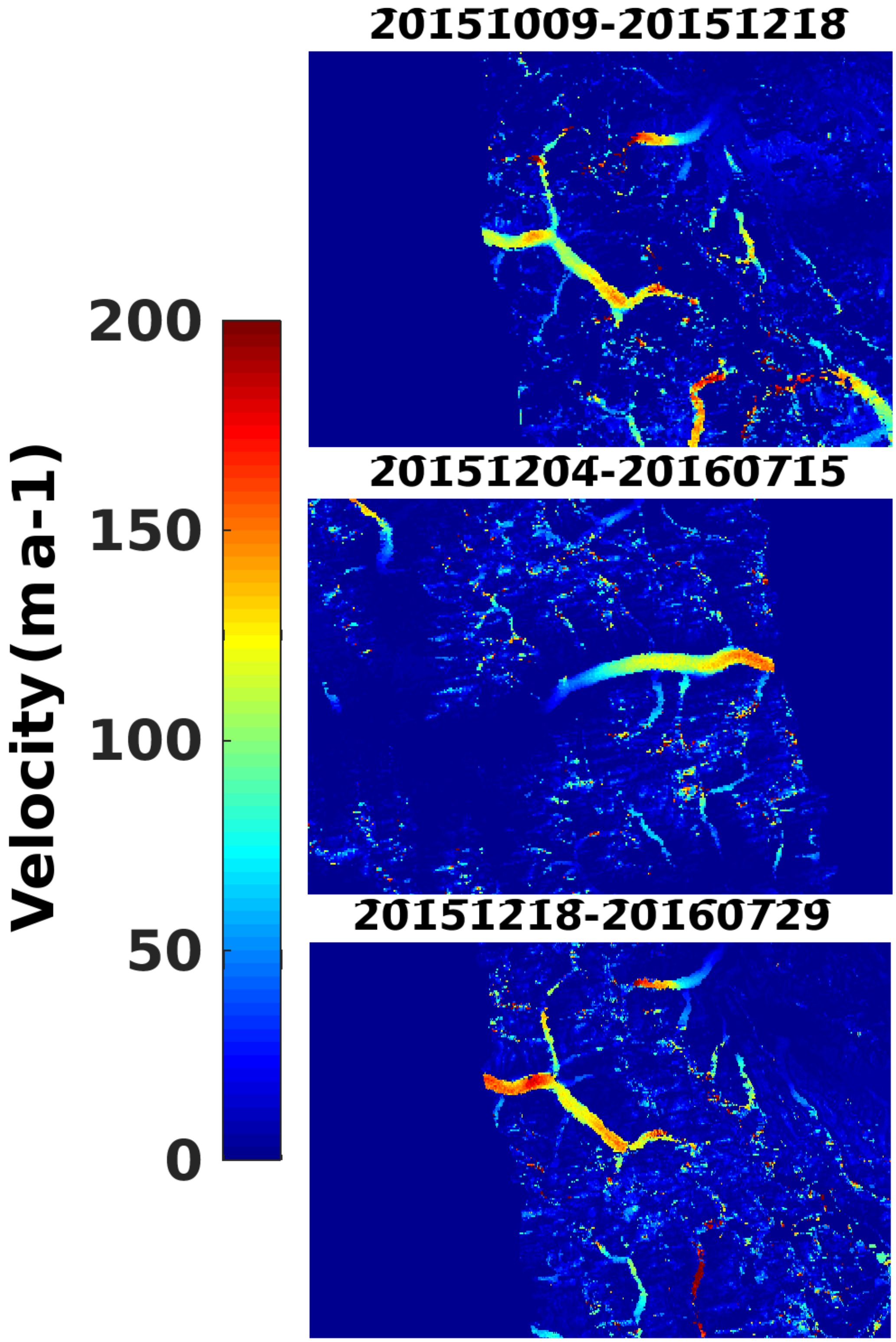
Velocity(m a-1)
200
150
100
50
0
20100215-20100703
20100703-20100818
20100818-20110103

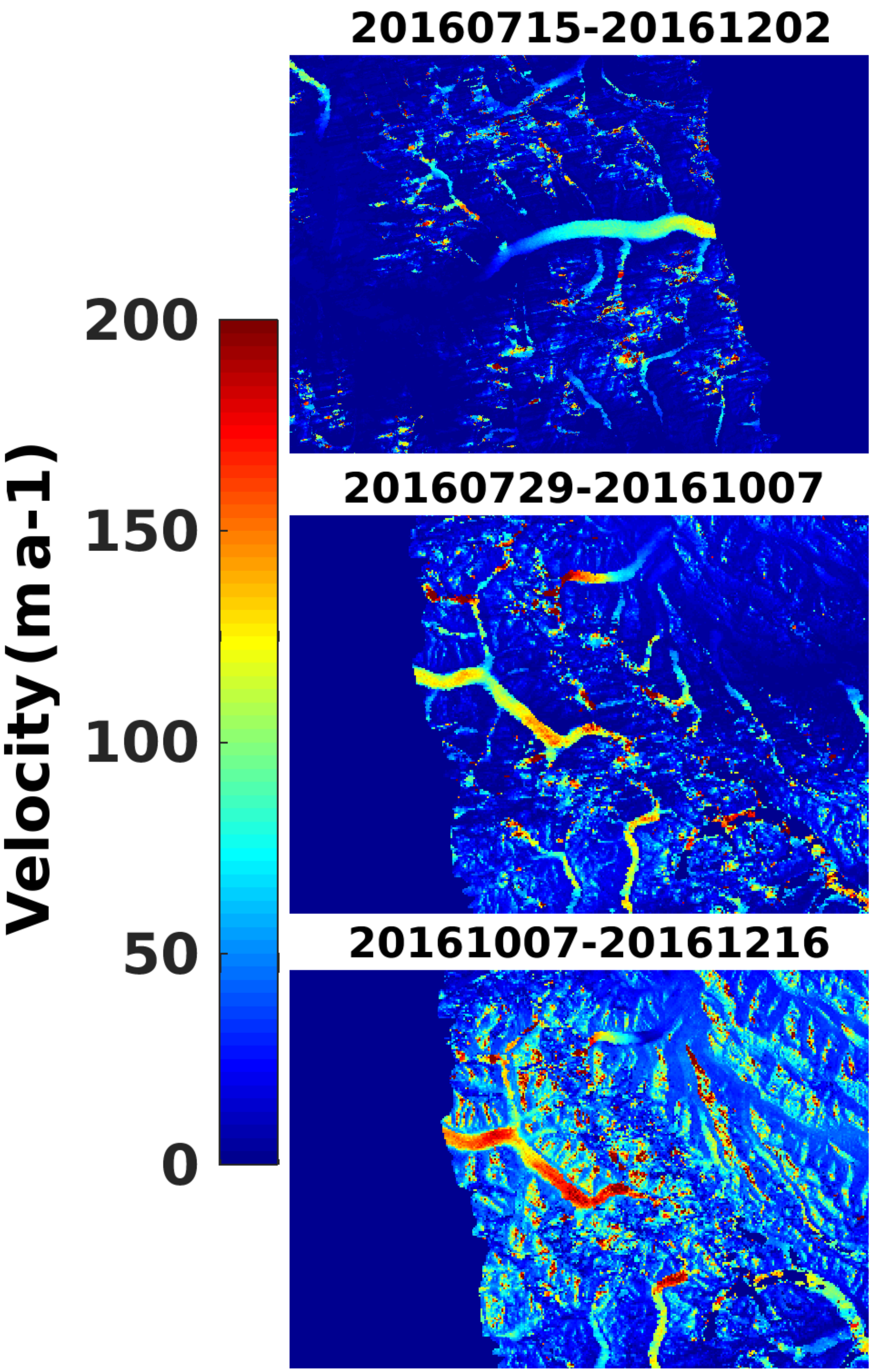


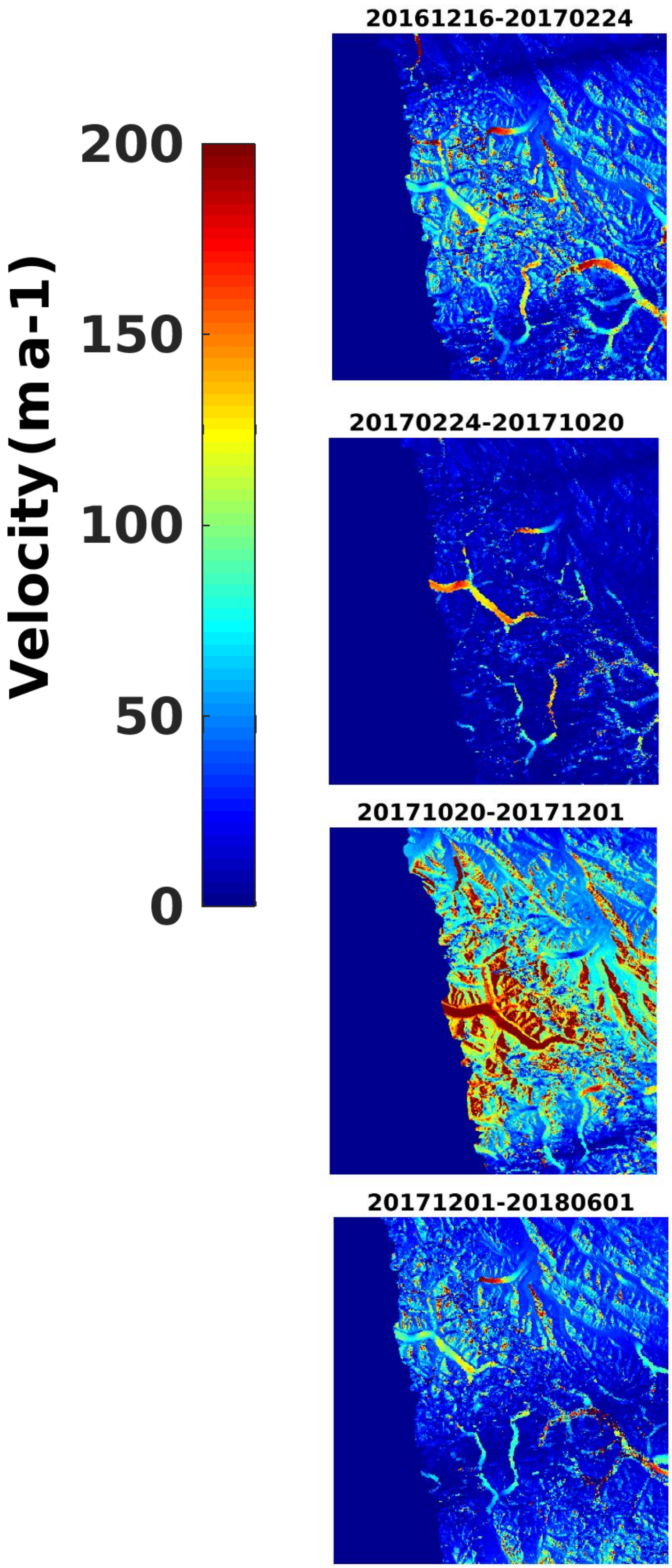
20110103-20110218
20140926-20150213
20141010-20141219
Velocity (m a-1)
200
150
100
50
0

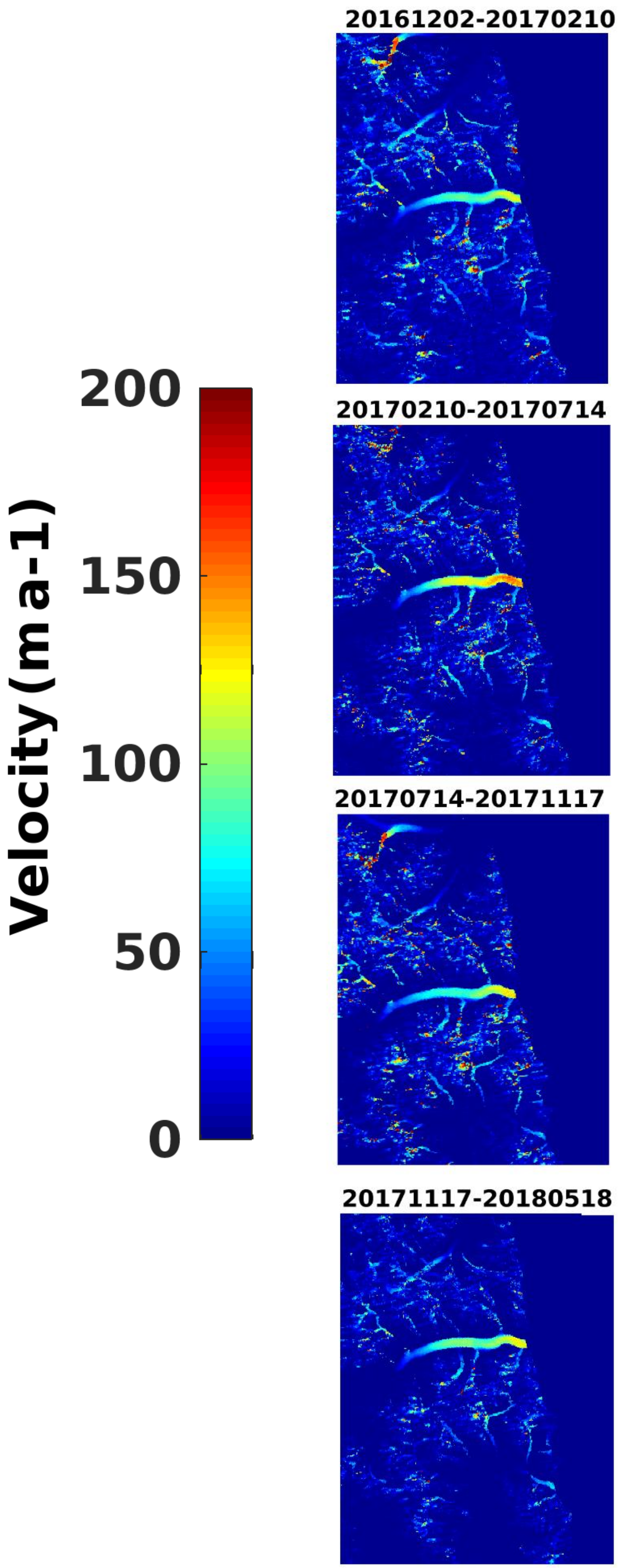

20141219-20150227
20150213-20150313
20150227-20150731
Velocity(m a-1)
200
150
100
50
0


20150313-20150925
20150731-20151009
20150925-20151204
200
150
100
50
0
Velocity(m a-1)

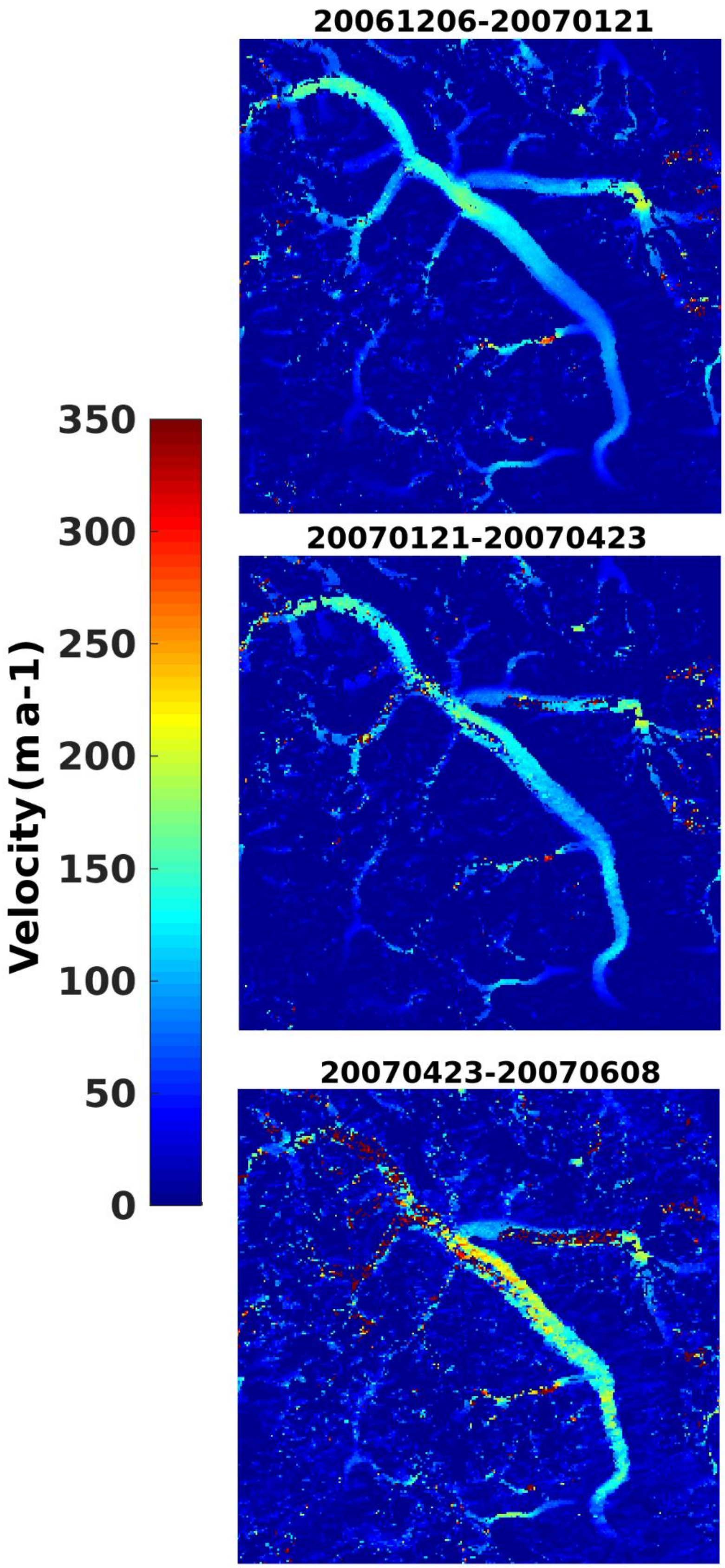

20151009-20151218
20151204-20160715
20151218-20160729
200
150
100
50
0
Velocity (m a-1)

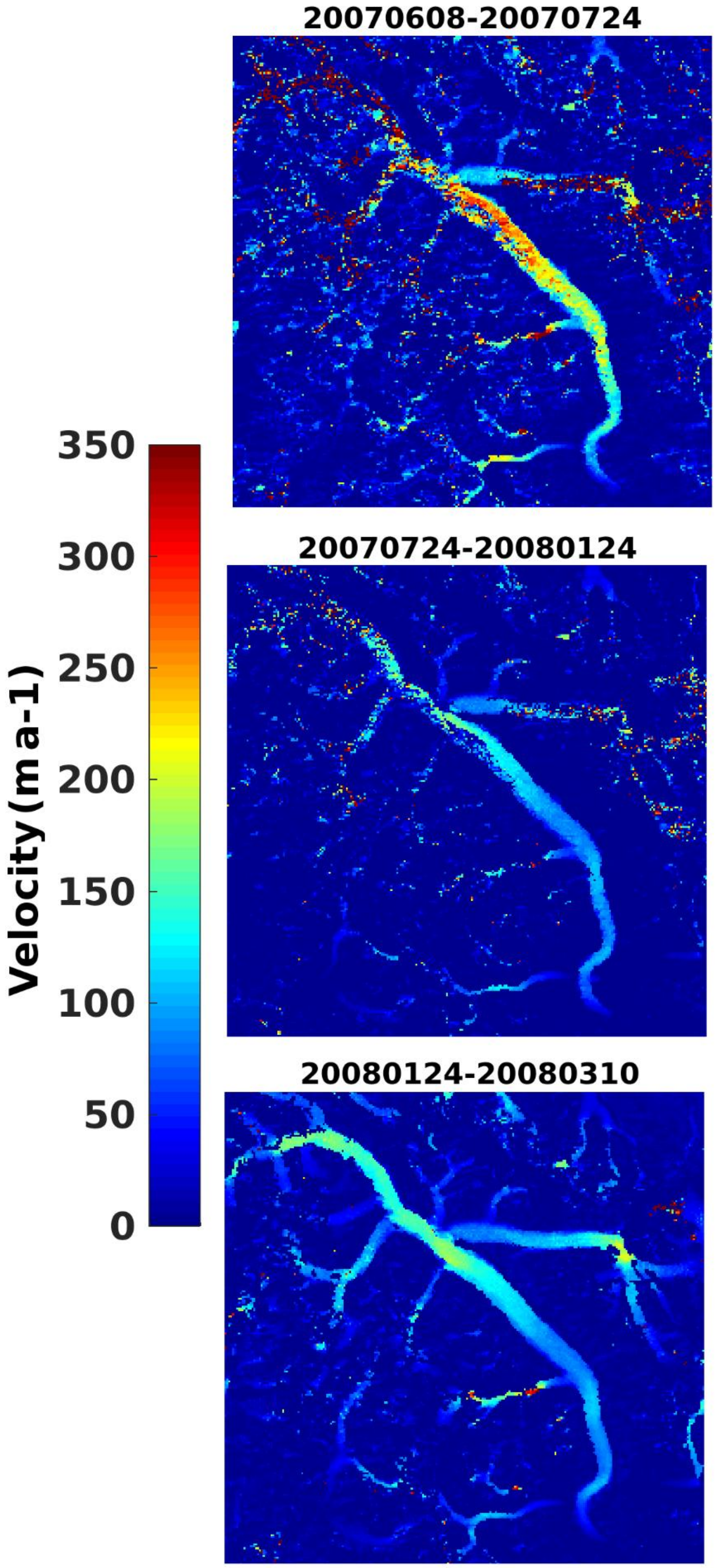

20160715-20161202
20160729-20161007
20161007-20161216
200
150
100
50
0
Velocity(m a-1)

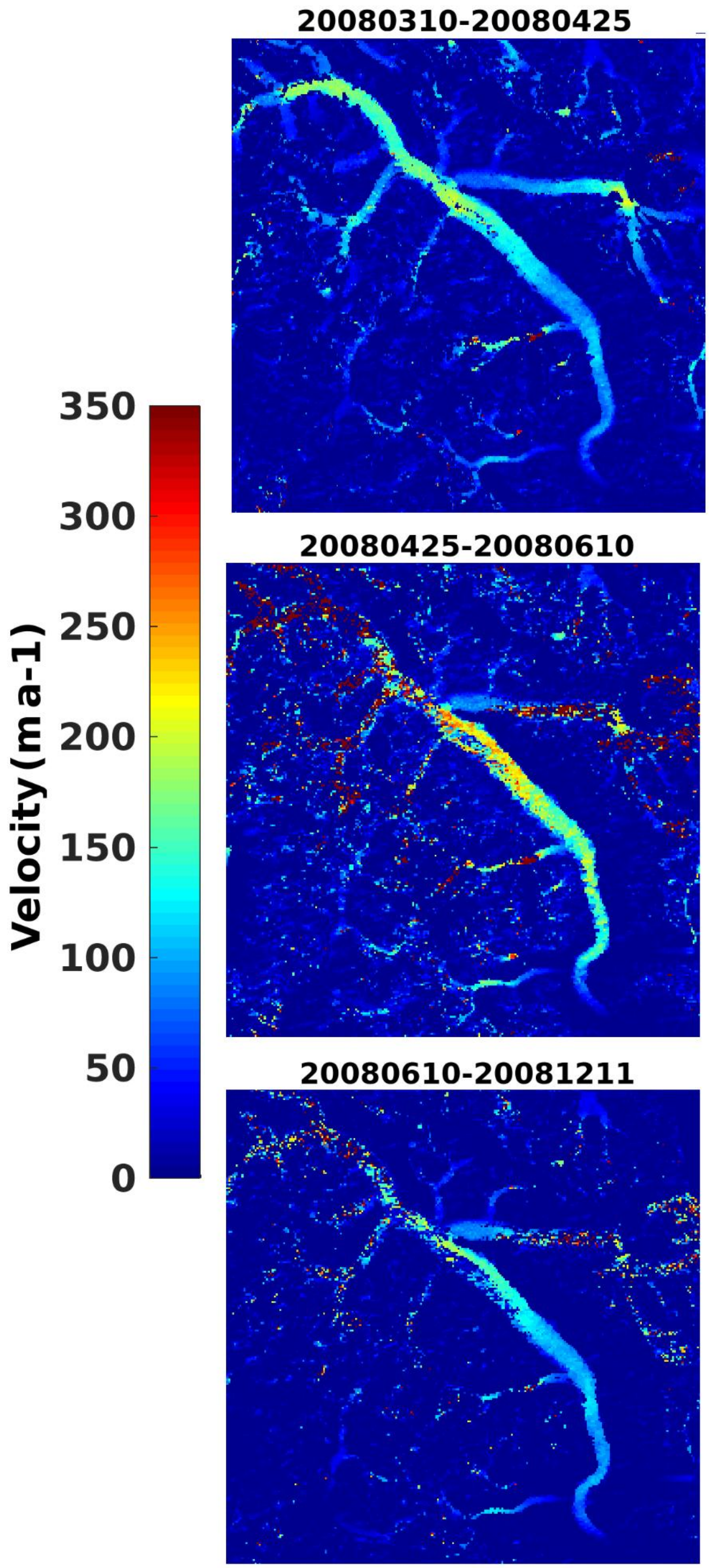

Velocity(m a-1)
200
150
100
50
0
20161216-20170224
20170224-20171020
20171020-20171201
20171201-20180601

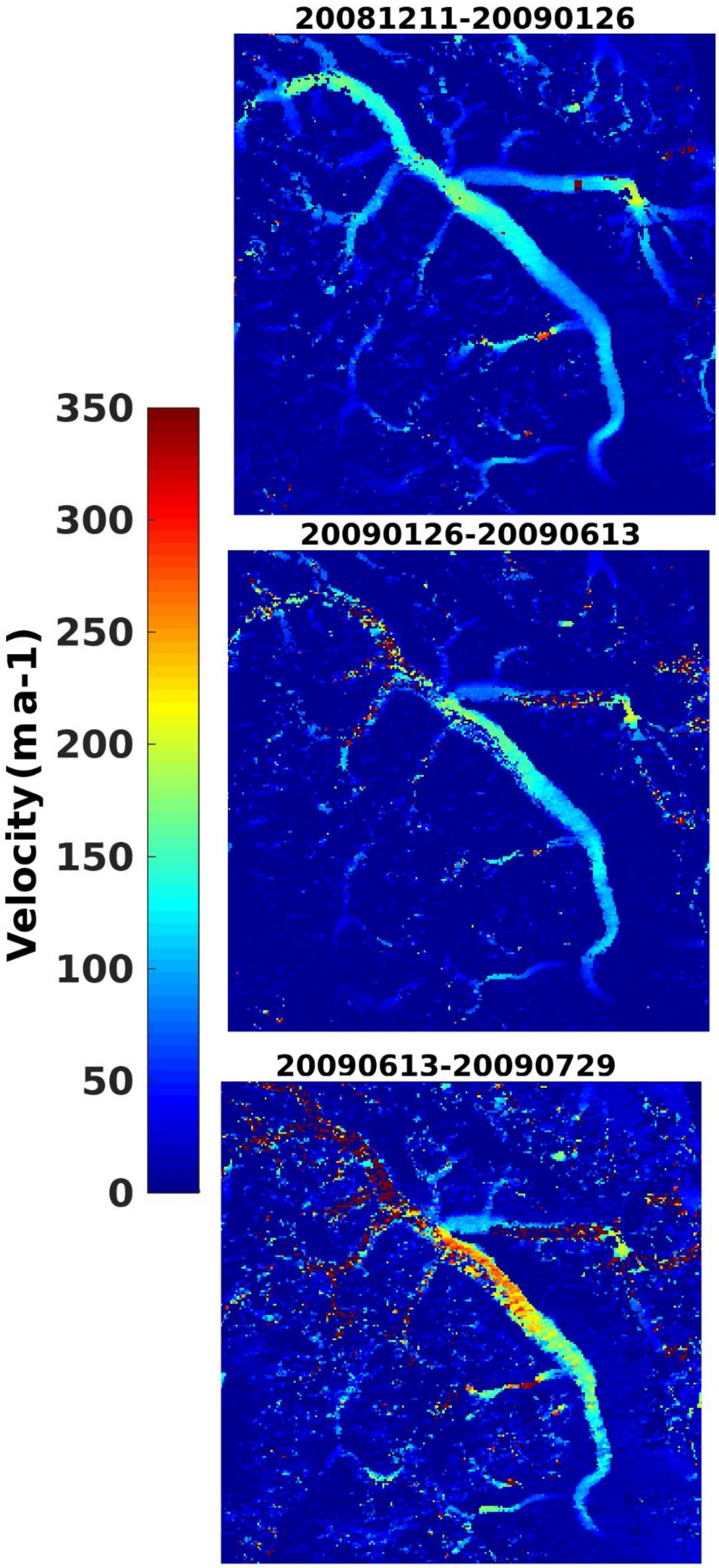

20161202-20170210
20170210-20170714
20170714-20171117
20171117-20180518
Velocity (m a-1)
200
150
100
50
0

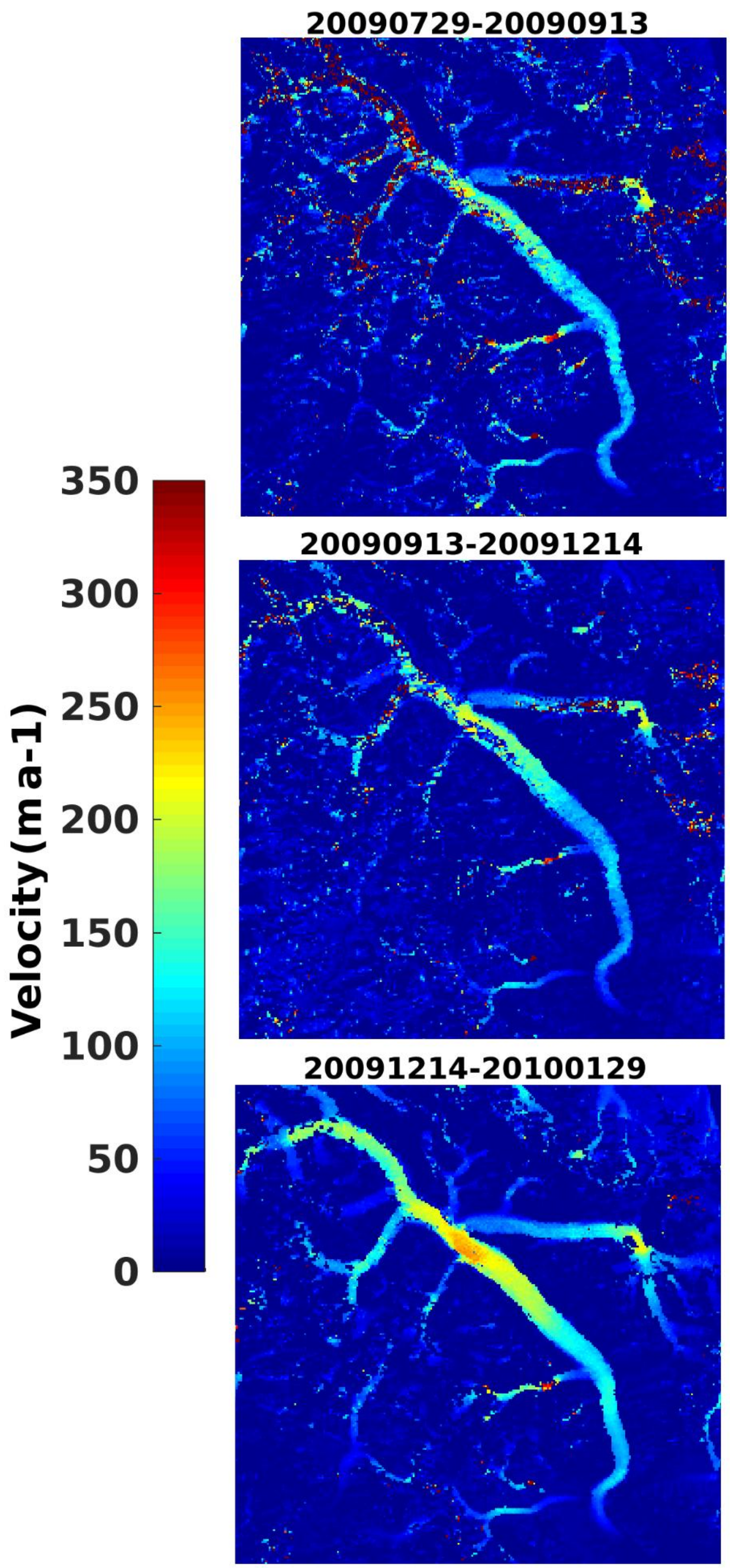
Surface velocity field of the Siachen Glacier during the study period (Figures S2)

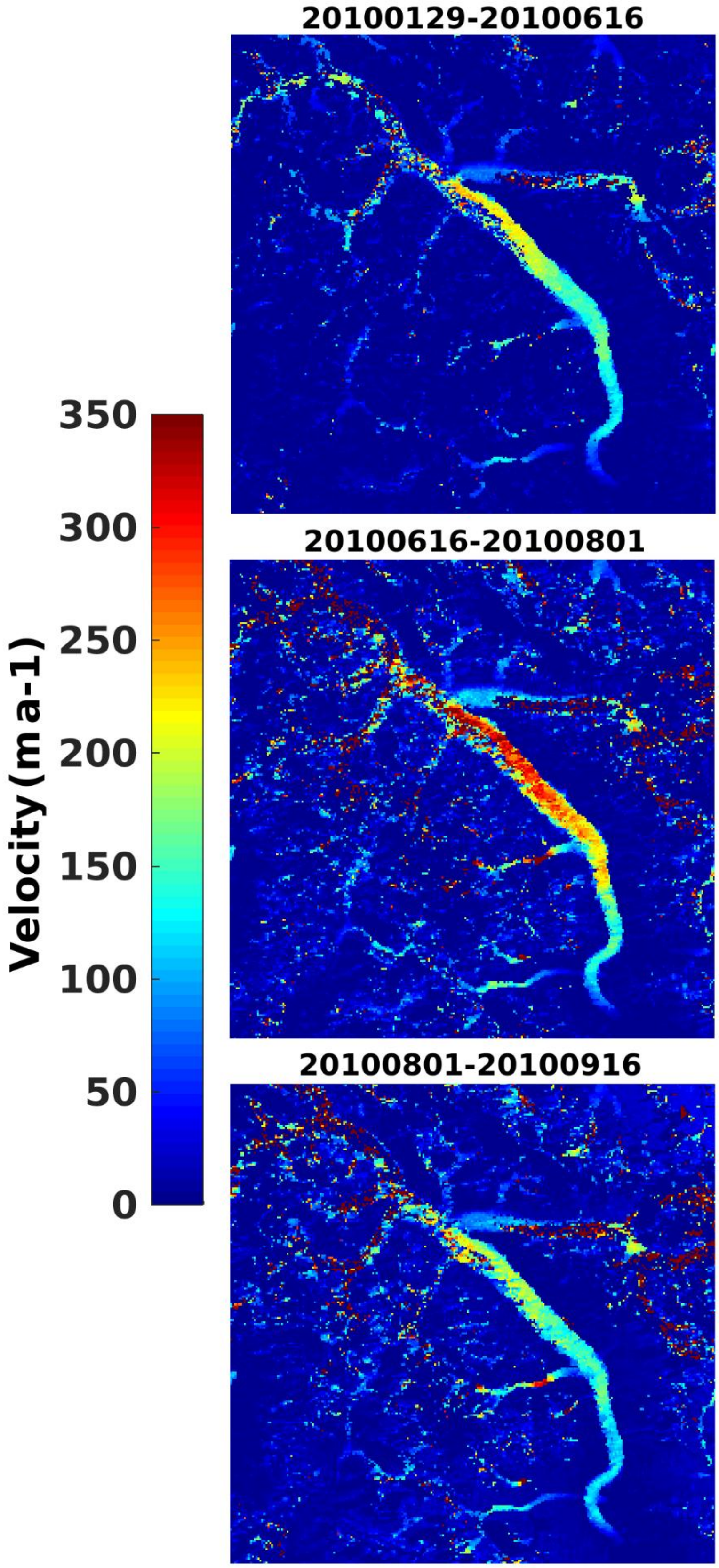

20061206-20070121
20070121-20070423
20070423-20070608
Velocity (m a-1)
350
300
250
200
150
100
50
0

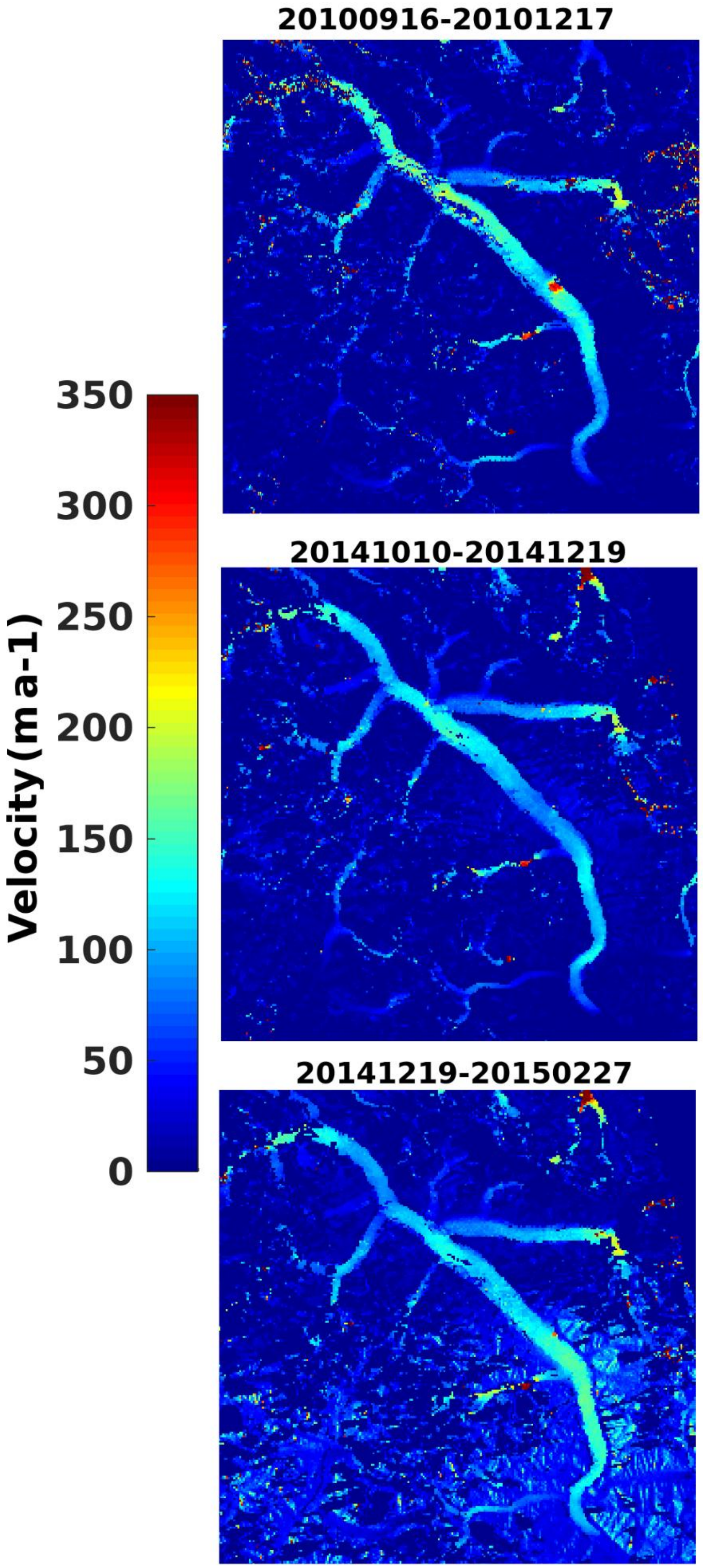

20070608-20070724
20070724-20080124
20080124-20080310
Velocity (m a-1)
350
300
250
200
150
100
50
0

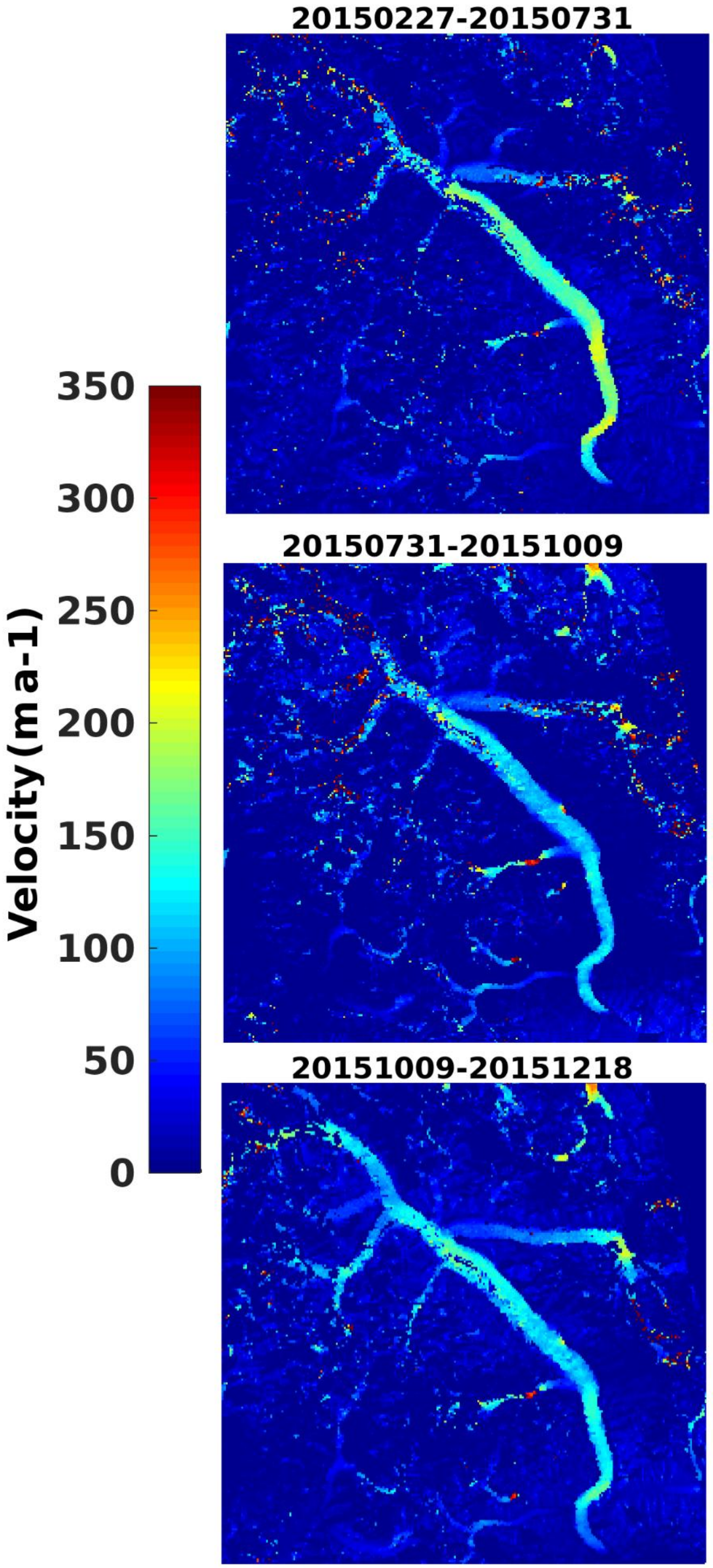

20080310-20080425
20080425-20080610
20080610-20081211
Velocity (m a-1)
350
300
250
200
150
100
50
0

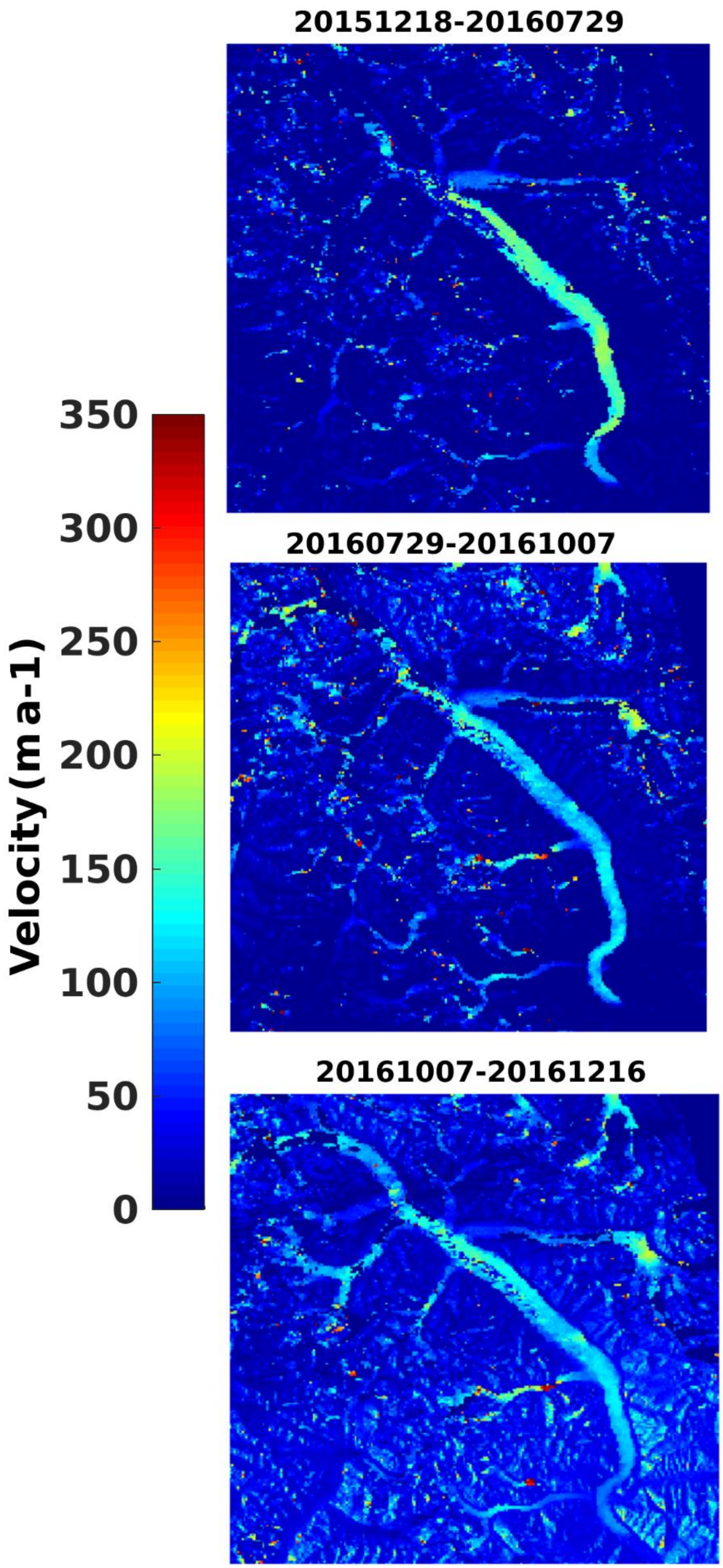

20081211-20090126
20090126-20090613
20090613-20090729
Velocity (m a-1)
350
300
250
200
150
100
50
0

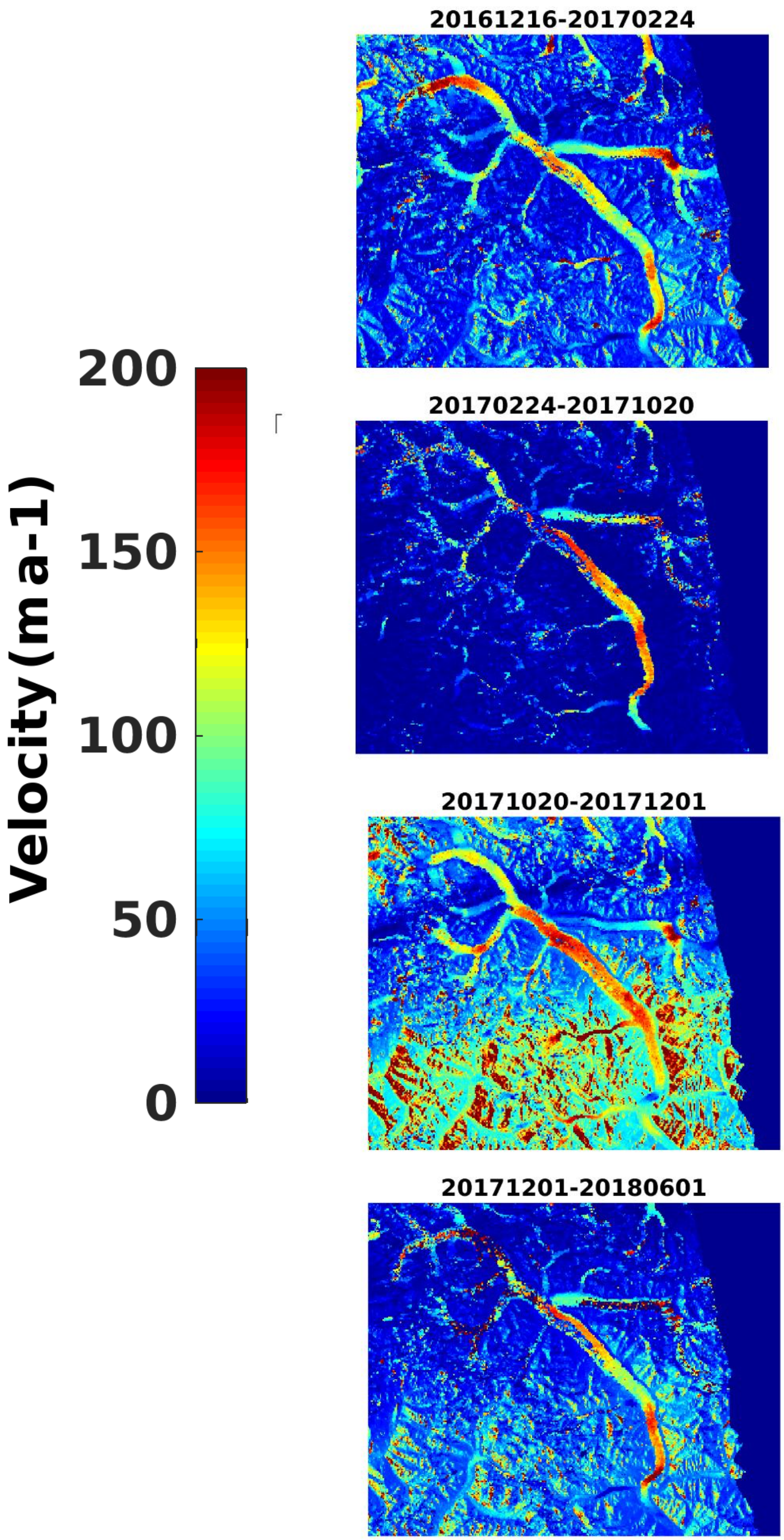

20090729-20090913
20090913-20091214
20091214-20100129
Velocity (m a-1)
350
300
250
200
150
100
50
0


20100129-20100616
20100616-20100801
20100801-20100916
Velocity (m a-1)
350
300
250
200
150
100
50
0


20100916-20101217
20141010-20141219
20141219-20150227
Velocity (m a-1)
350
300
250
200
150
100
50
0

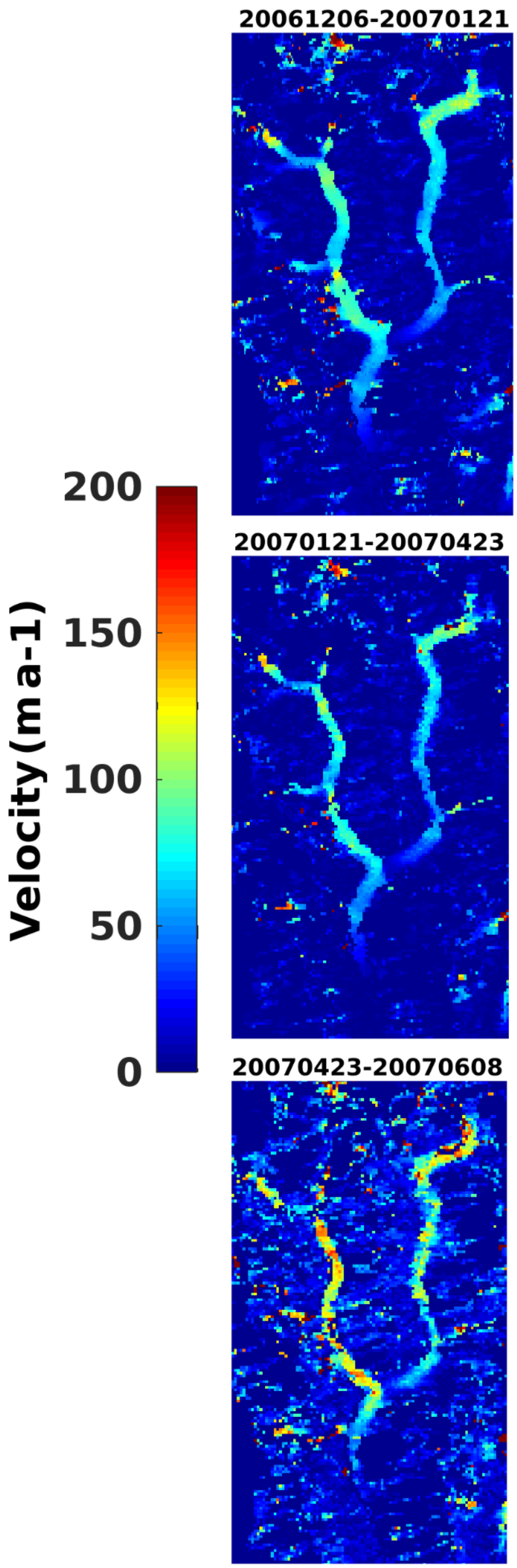

20150227-20150731
20150731-20151009
20151009-20151218
350
300
250
200
150
100
50
0
Velocity (m a-1)

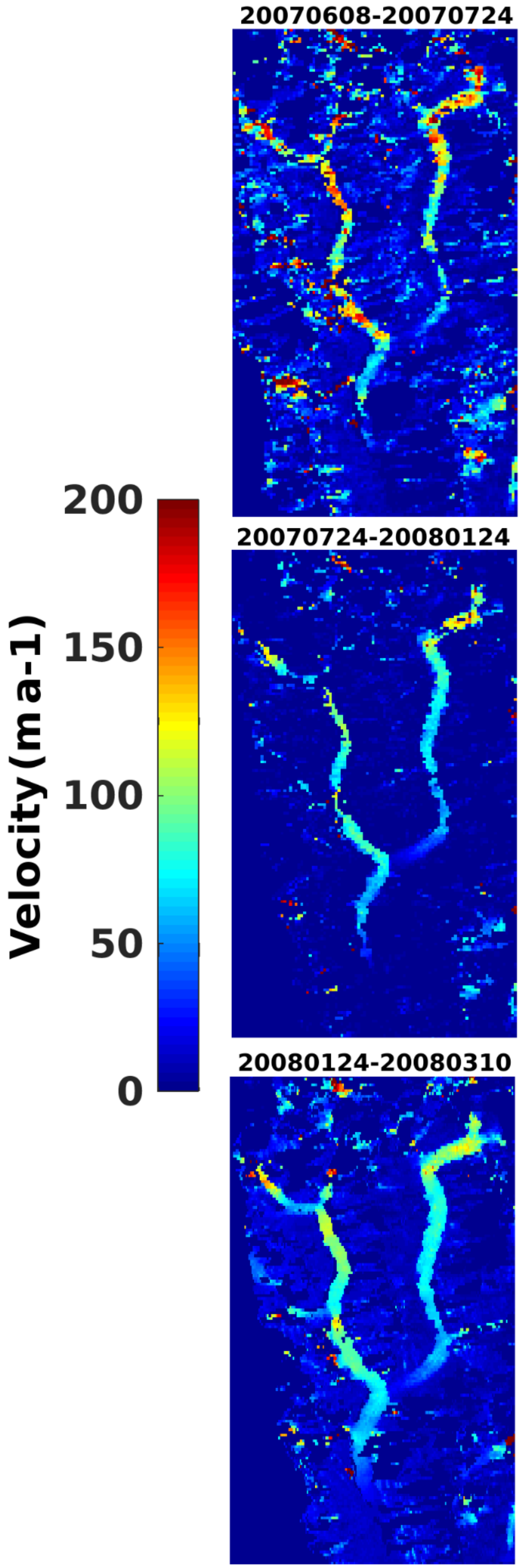

20151218-20160729
20160729-20161007
20161007-20161216
Velocity (m a-1)
350
300
250
200
150
100
50
0

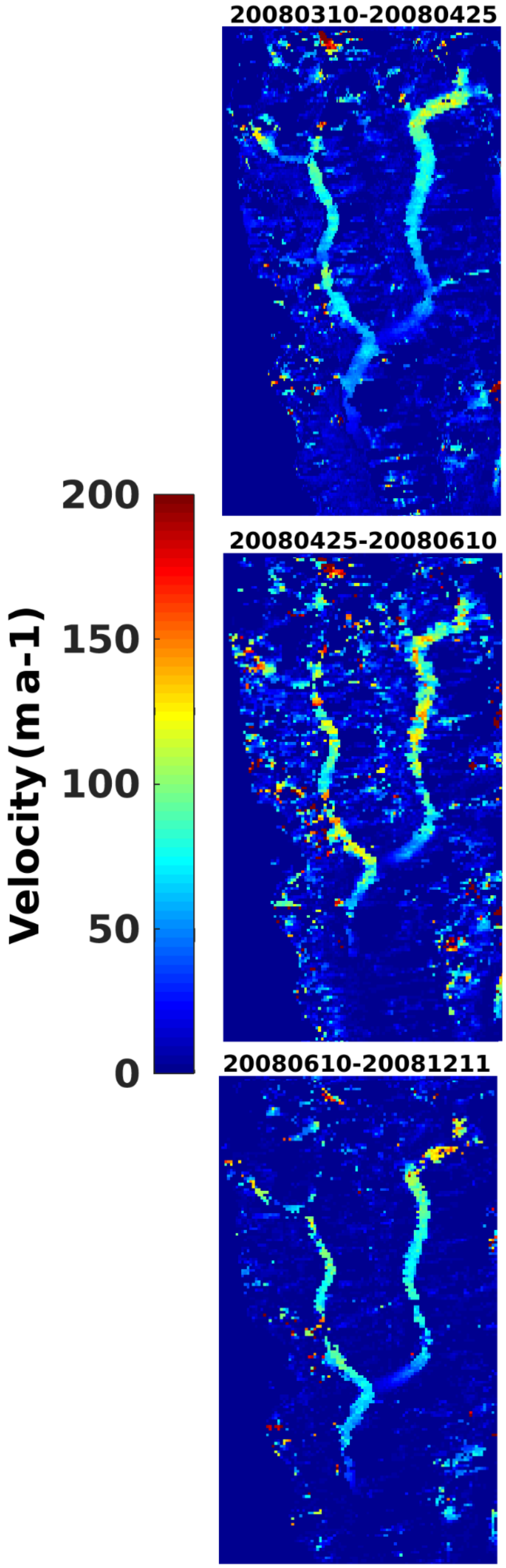


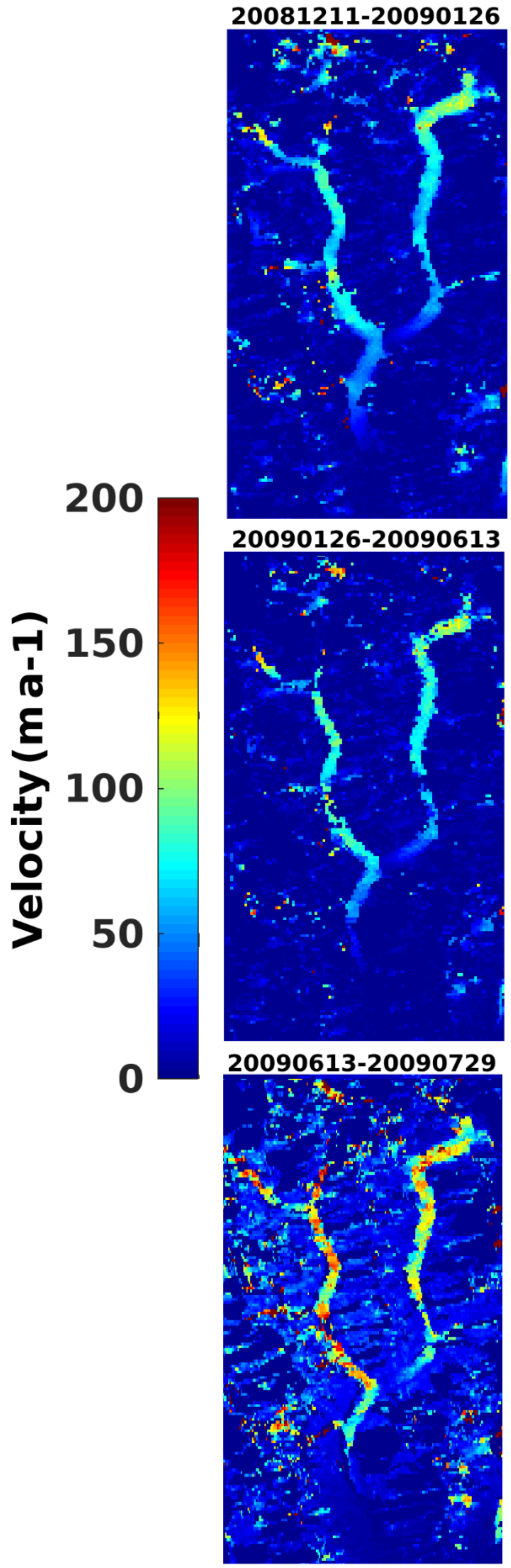
20161216-20170224
200
150
100
50
0
Velocity(m a-1)
20170224-20171020
20171020-20171201
20171201-20180601

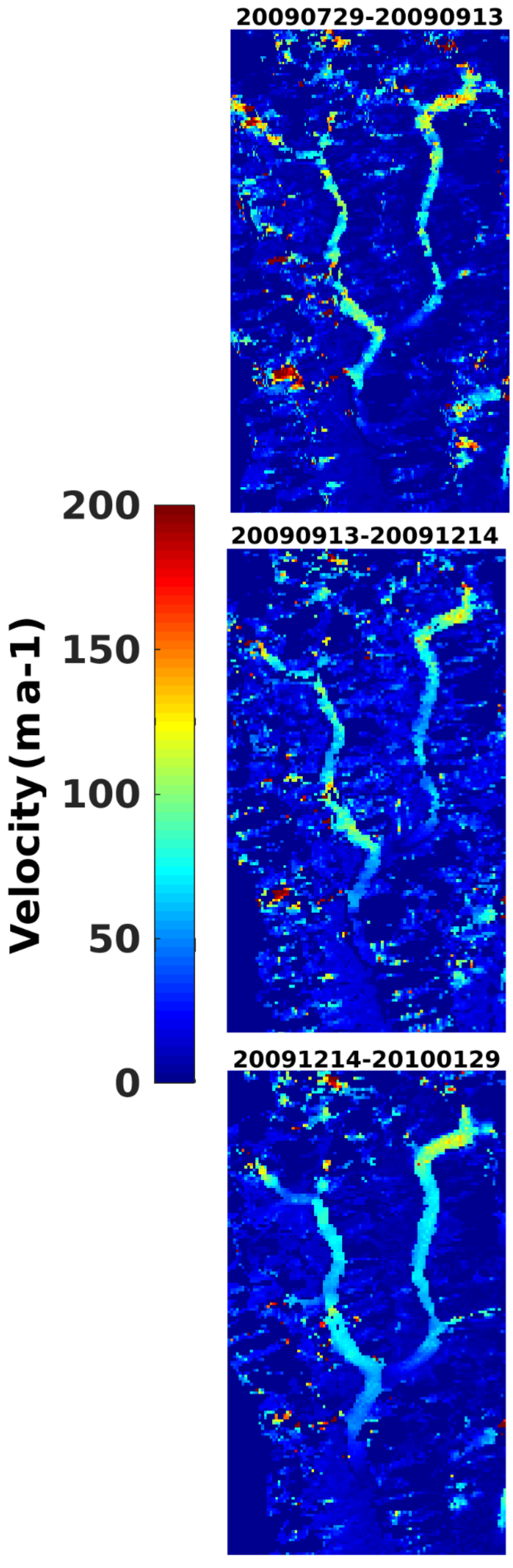
Surface velocity field of the Kundos Glacier during the study period (Figures S3)

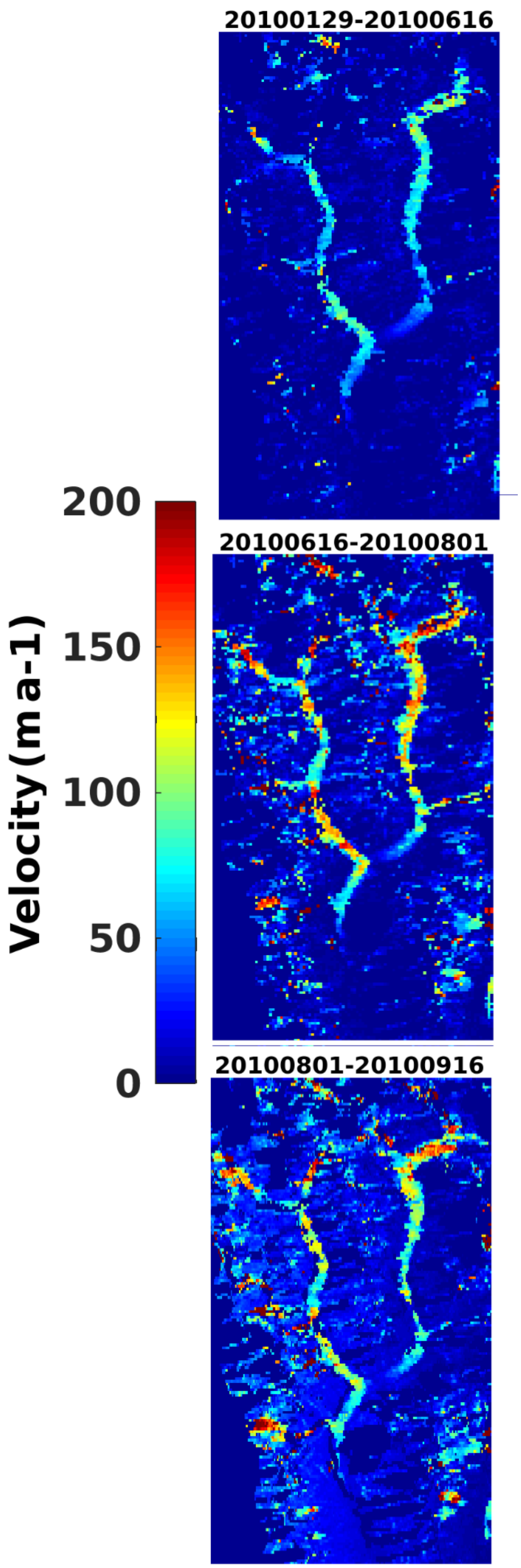

20061206-20070121
20070121-20070423
20070423-20070608
Velocity (m a-1)
200
150
100
50
0

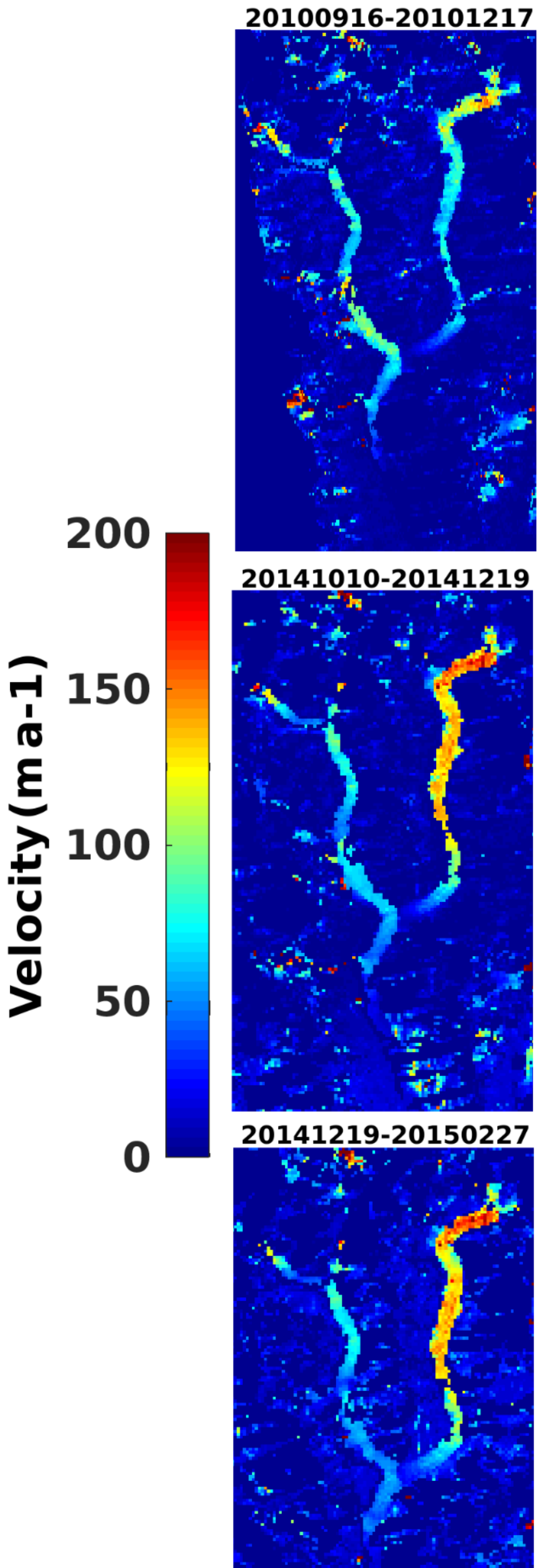

20070608-20070724
20070724-20080124
20080124-20080310
200
150
100
50
0
Velocity (m a-1)


20080310-20080425
20080425-20080610
20080610-20081211
200
150
100
50
0
Velocity (m a-1)

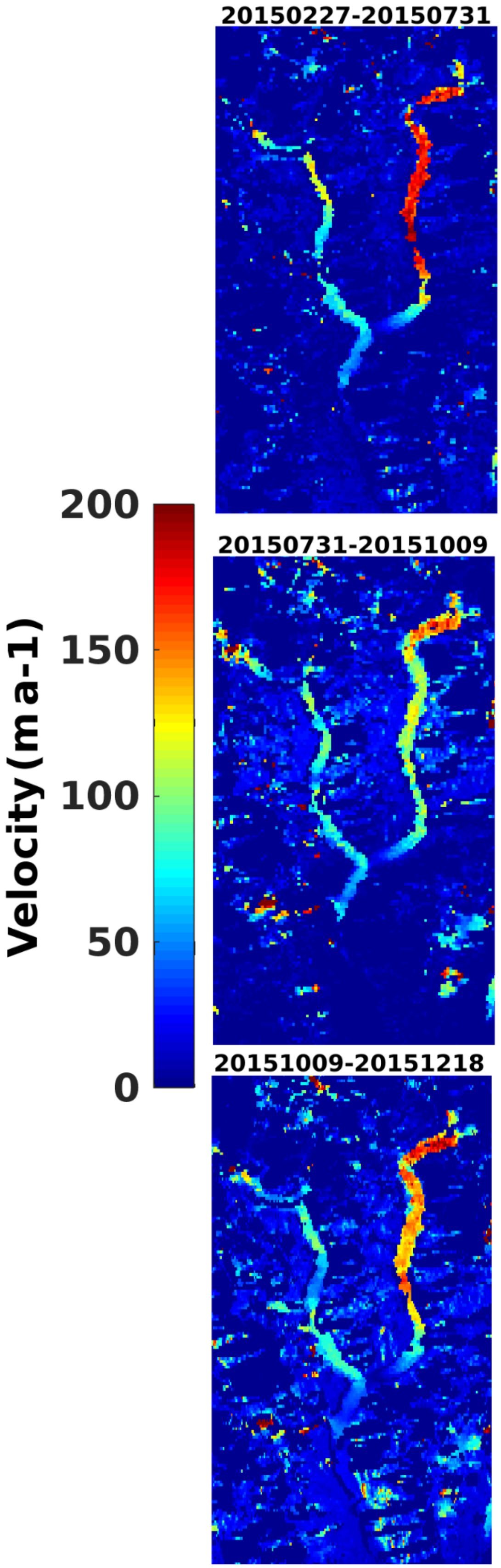

20081211-20090126
20090126-20090613
20090613-20090729
200
150
100
50
0
Velocity (m a-1)

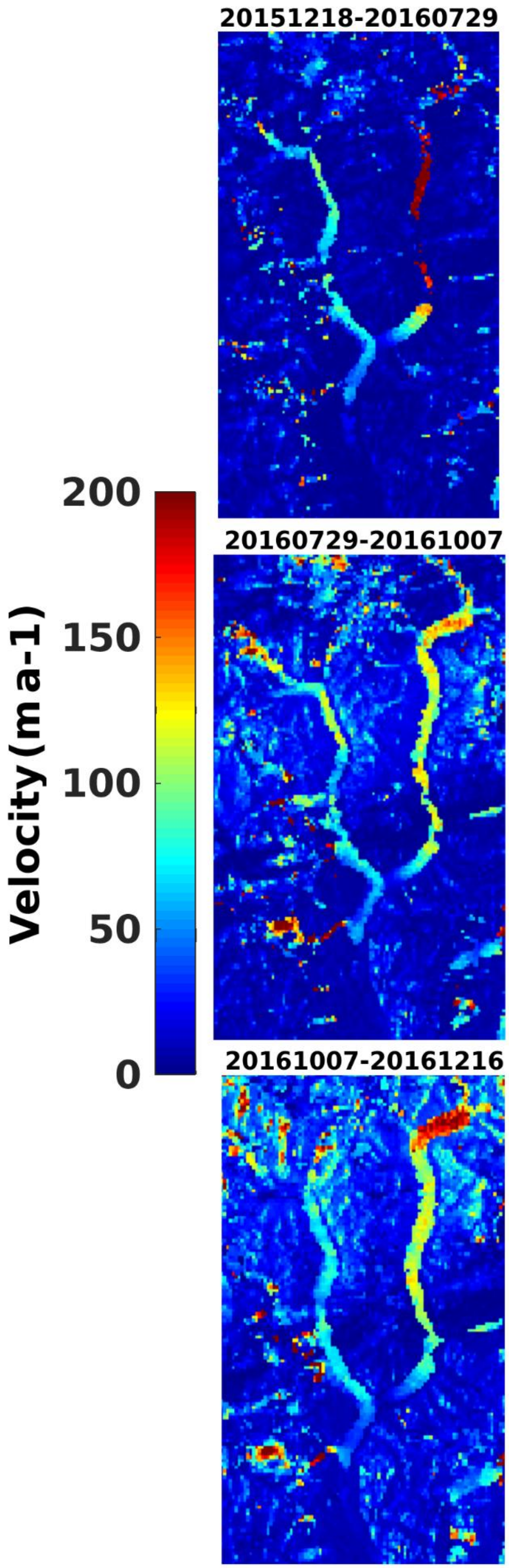

20090729-20090913
20090913-20091214
20091214-20100129
Velocity (m a-1)
200
150
100
50
0

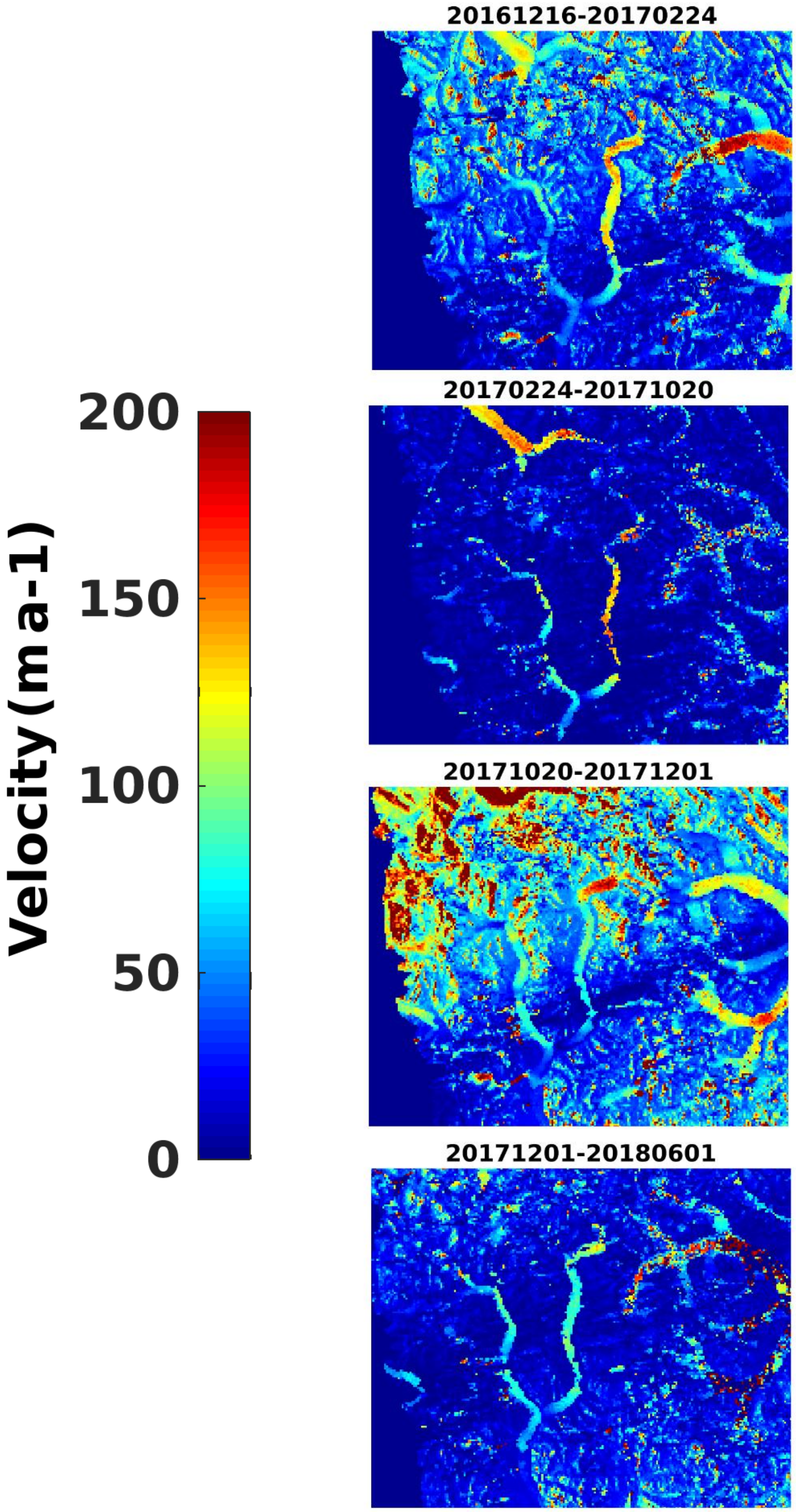

20100129-20100616
20100616-20100801
20100801-20100916
Velocity (m a-1)
200
150
100
50
0



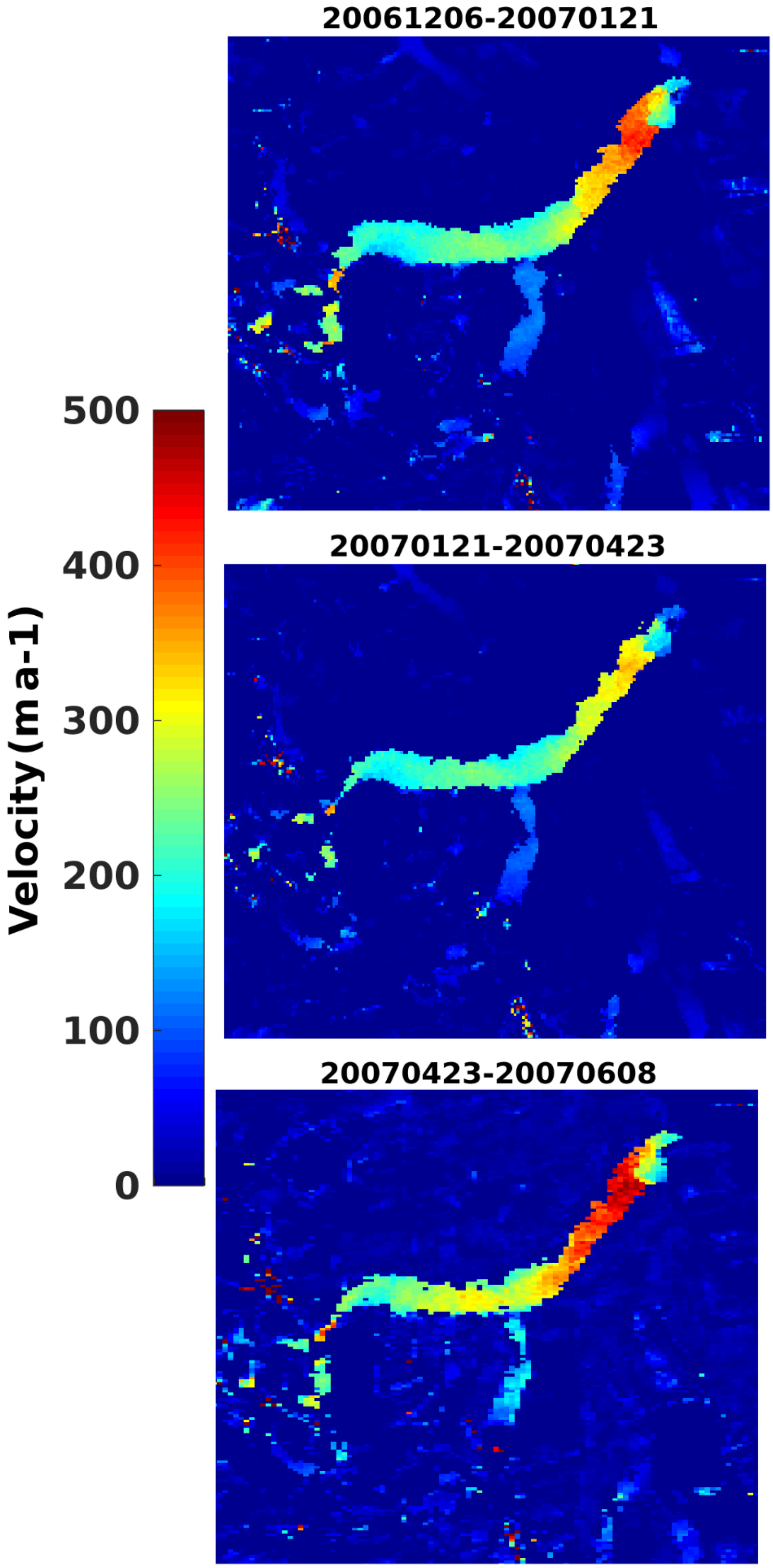
20100916-20101217
20141010-20141219
20141219-20150227
Velocity (m a-1)
200
150
100
50
0

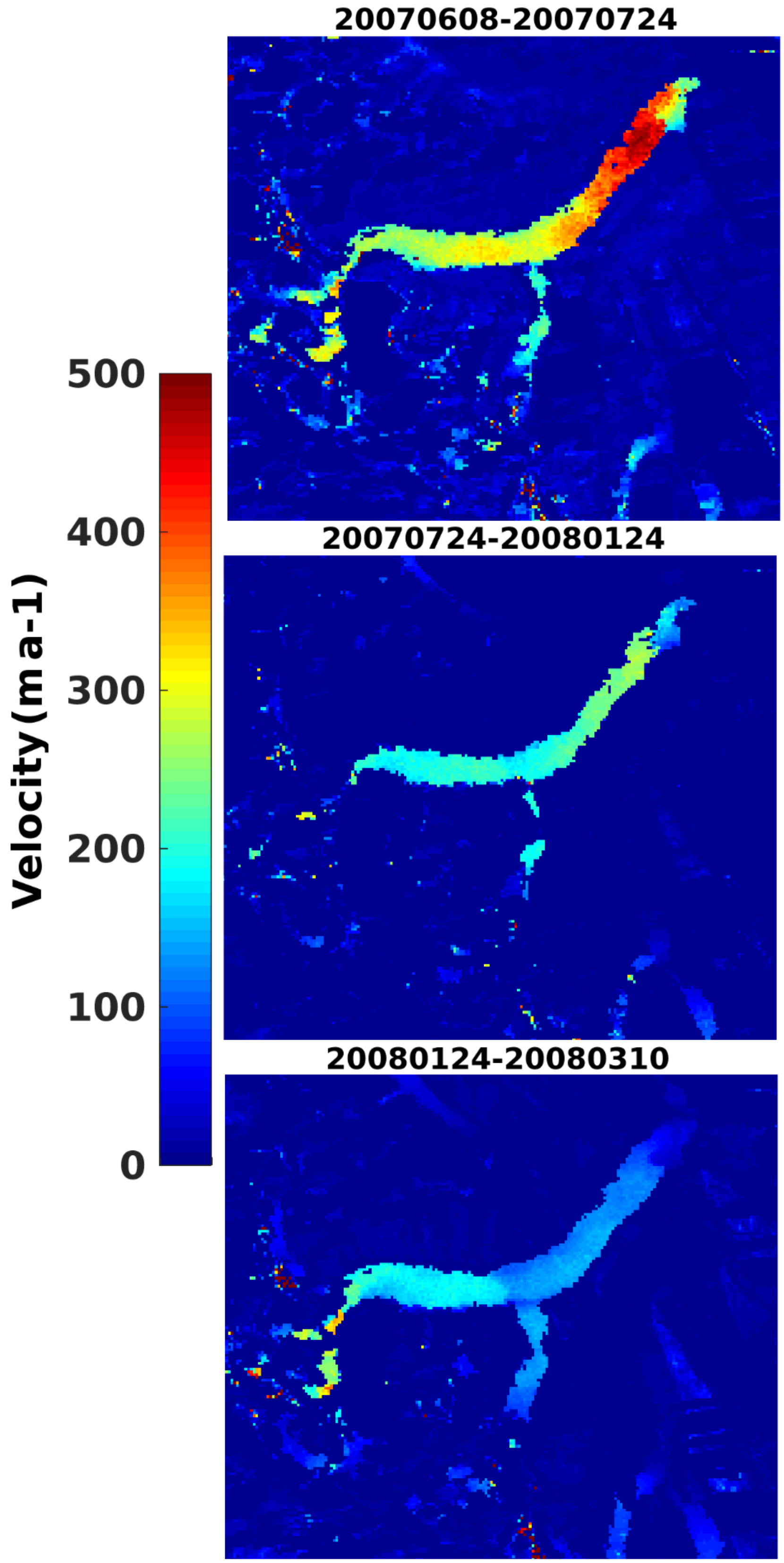

20150227-20150731
20150731-20151009
20151009-20151218
Velocity (m a-1)
200
150
100
50
0

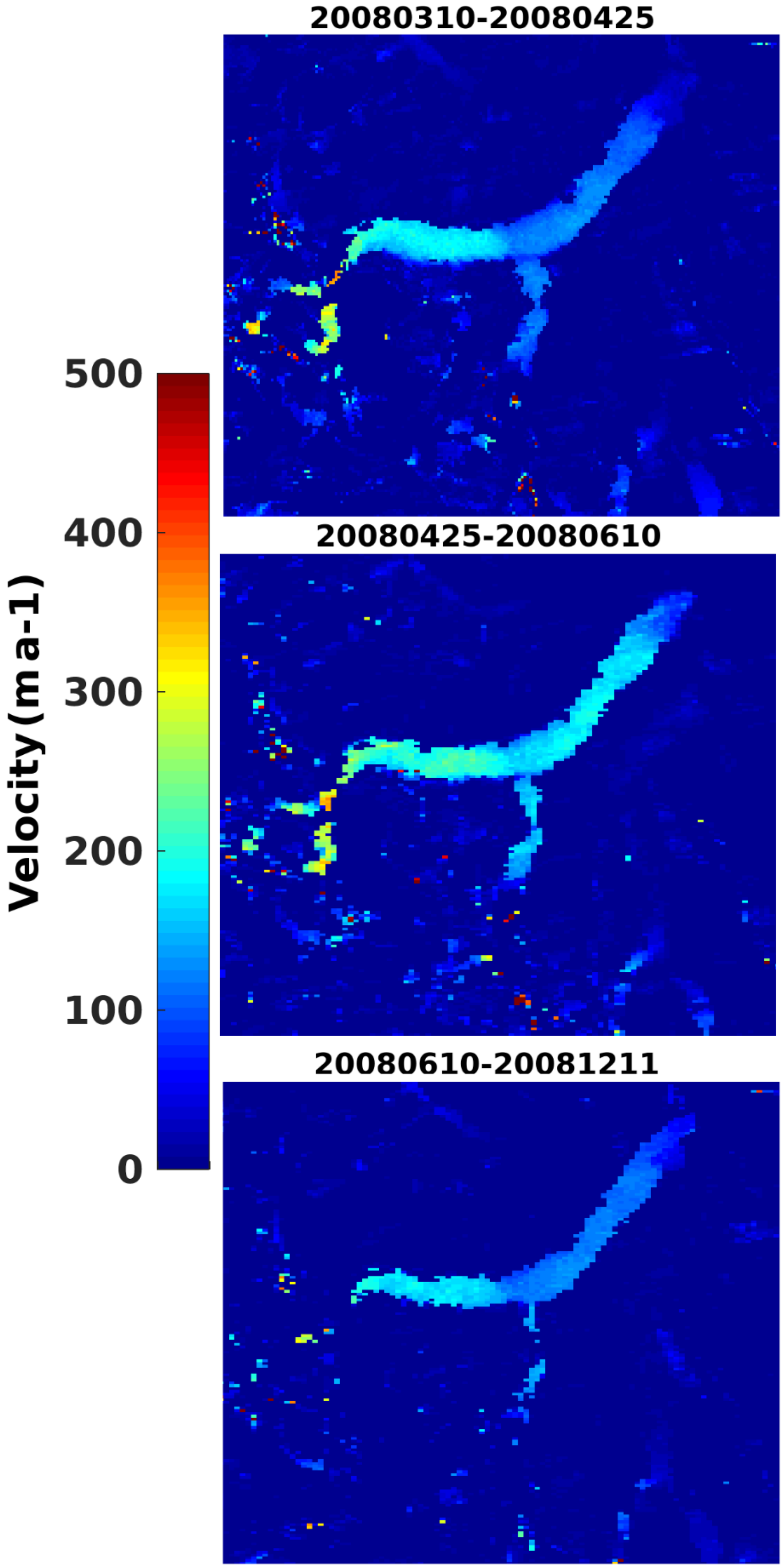

20151218-20160729
20160729-20161007
20161007-20161216
Velocity (m a-1)
200
150
100
50
0


20161216-20170224
20170224-20171020
20171020-20171201
20171201-20180601
Velocity (m a-1)
200
150
100
50
0

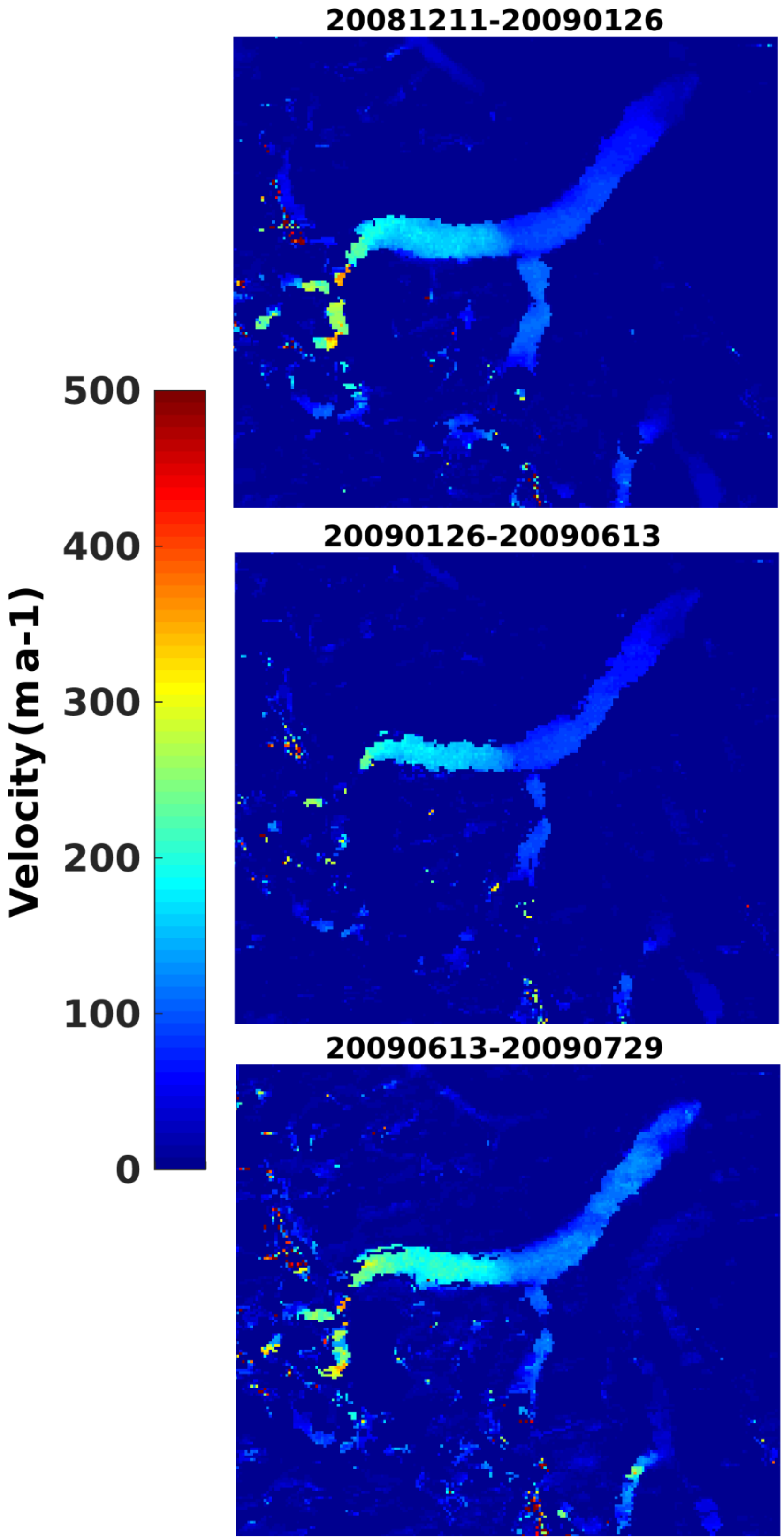
Surface velocity field of Gasherbrum Glacier during the study period (Figures S4)

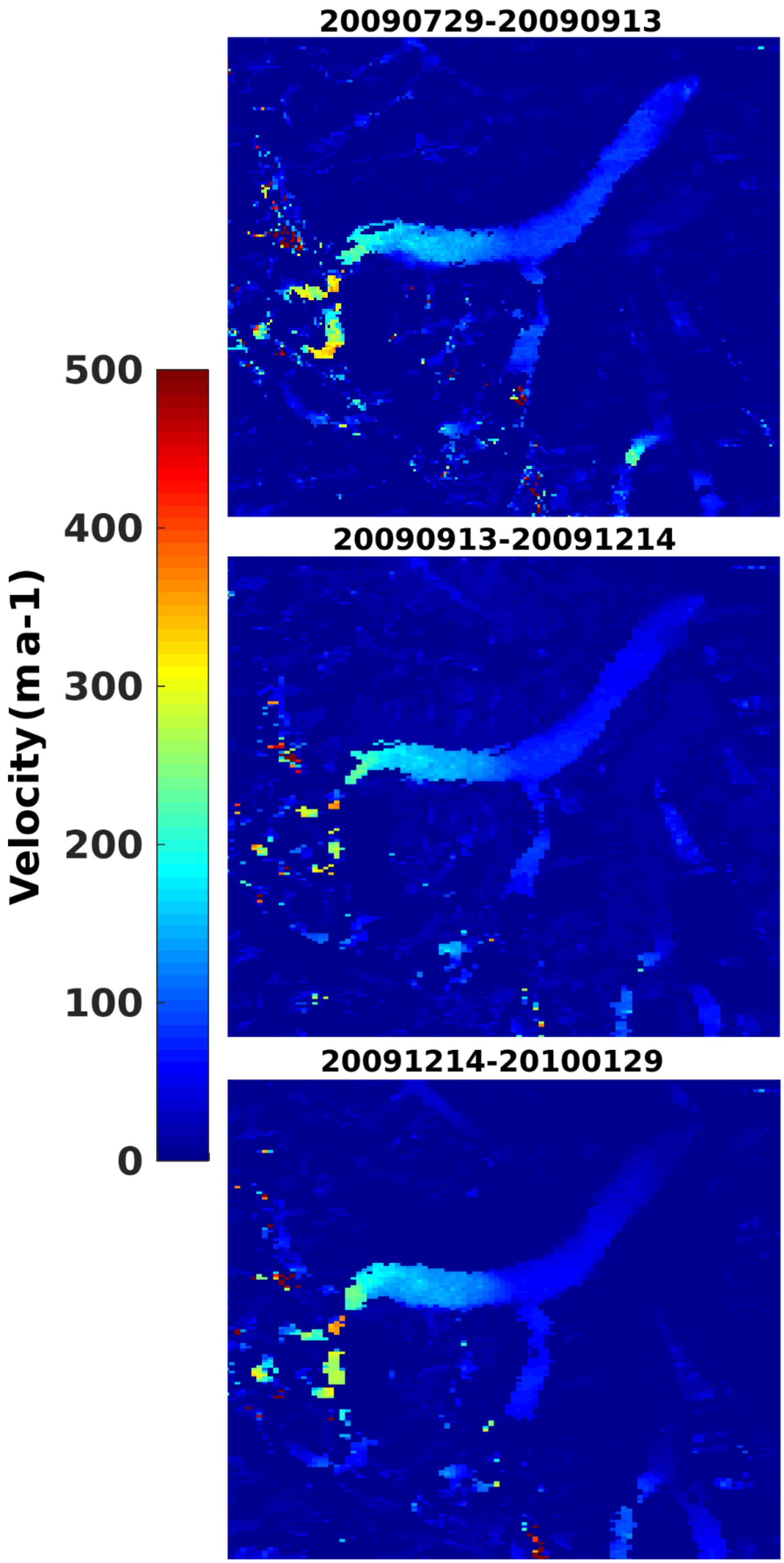

20061206-20070121
20070121-20070423
20070423-20070608
Velocity (m a-1)
500
400
300
200
100
0

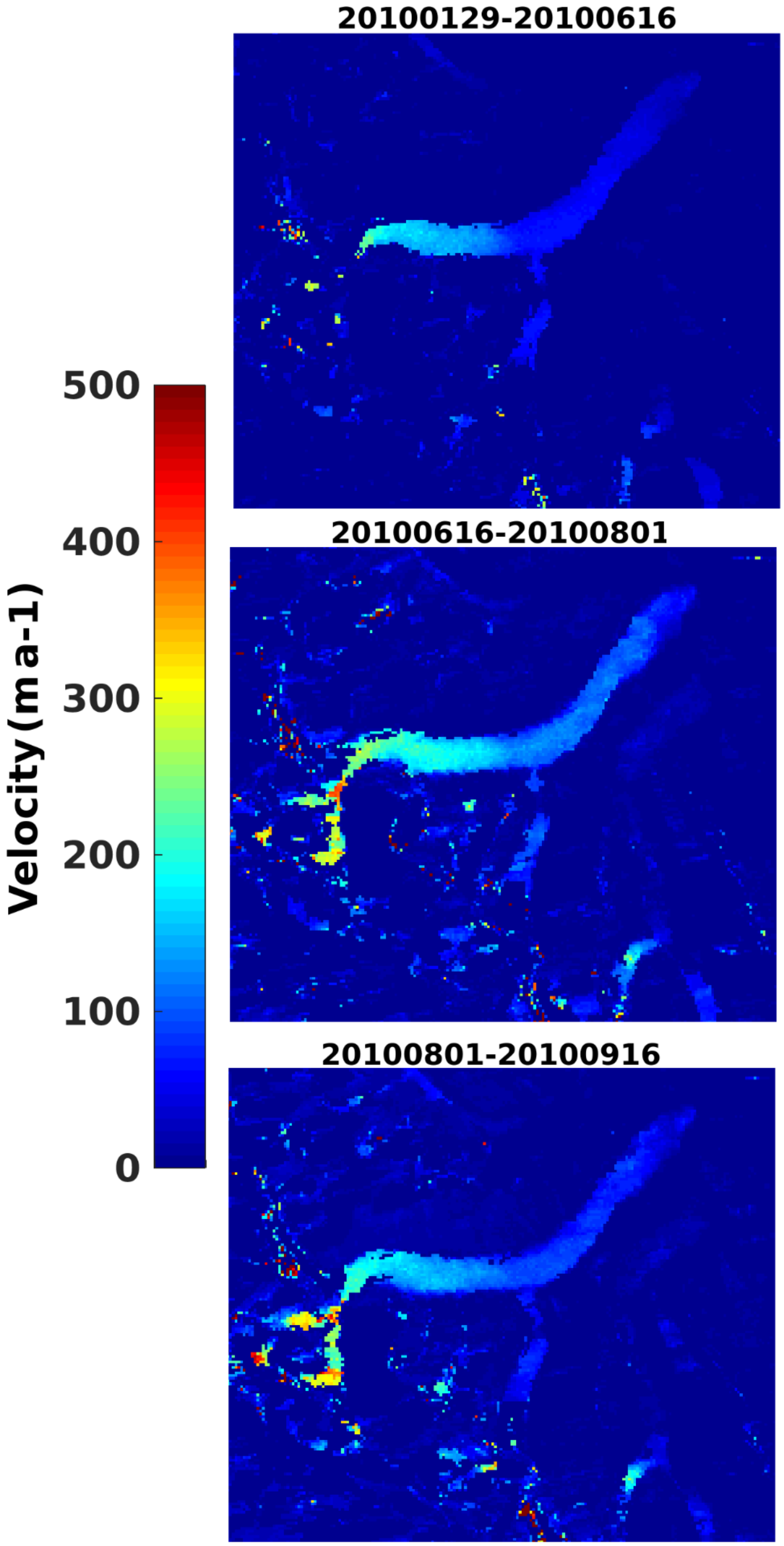

20070608-20070724
20070724-20080124
20080124-20080310
500
400
300
200
100
0
Velocity (m a-1)

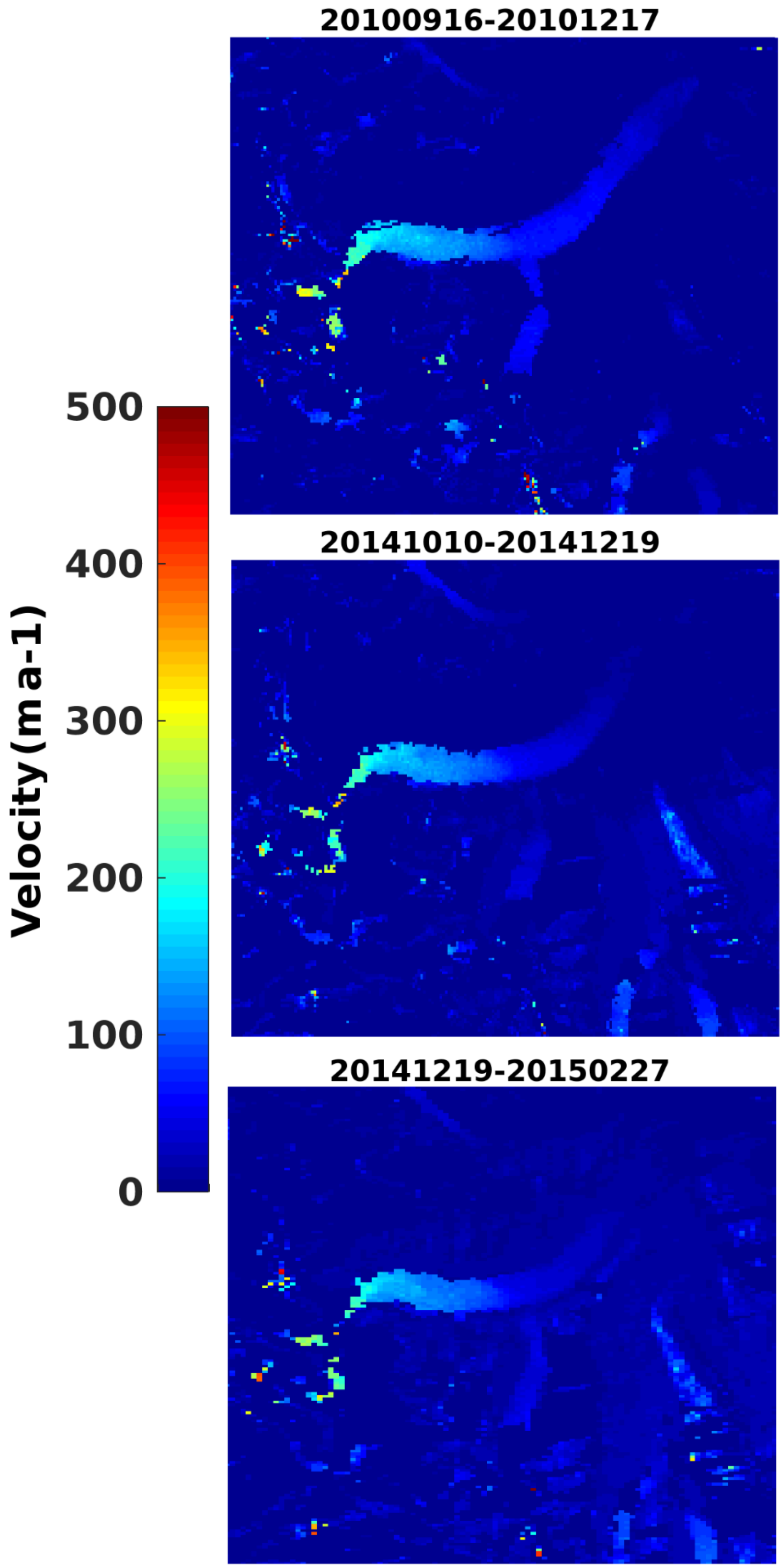


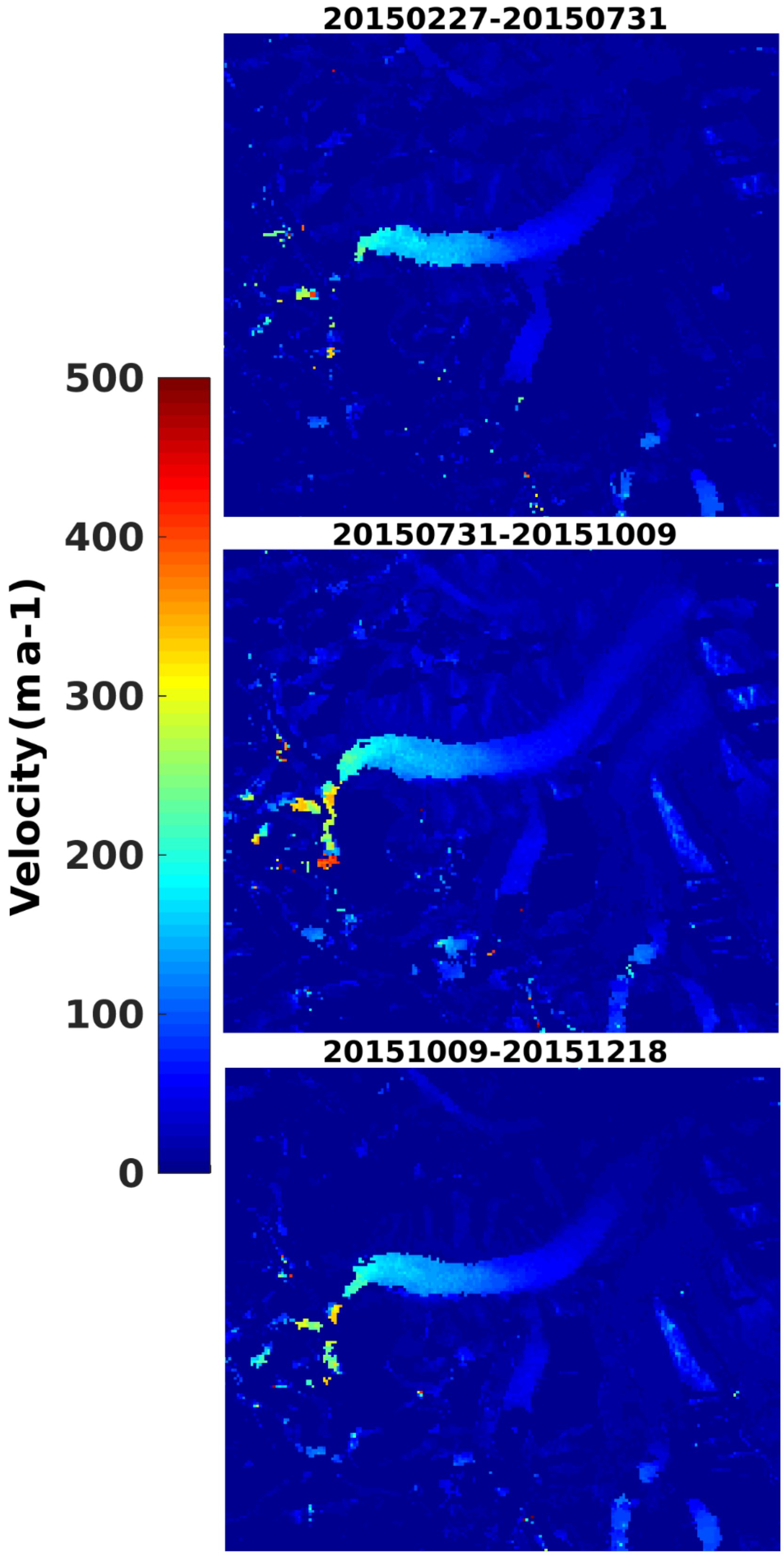
20080310-20080425
20080425-20080610
20080610-20081211
Velocity (m a-1)
500
400
300
200
100
0

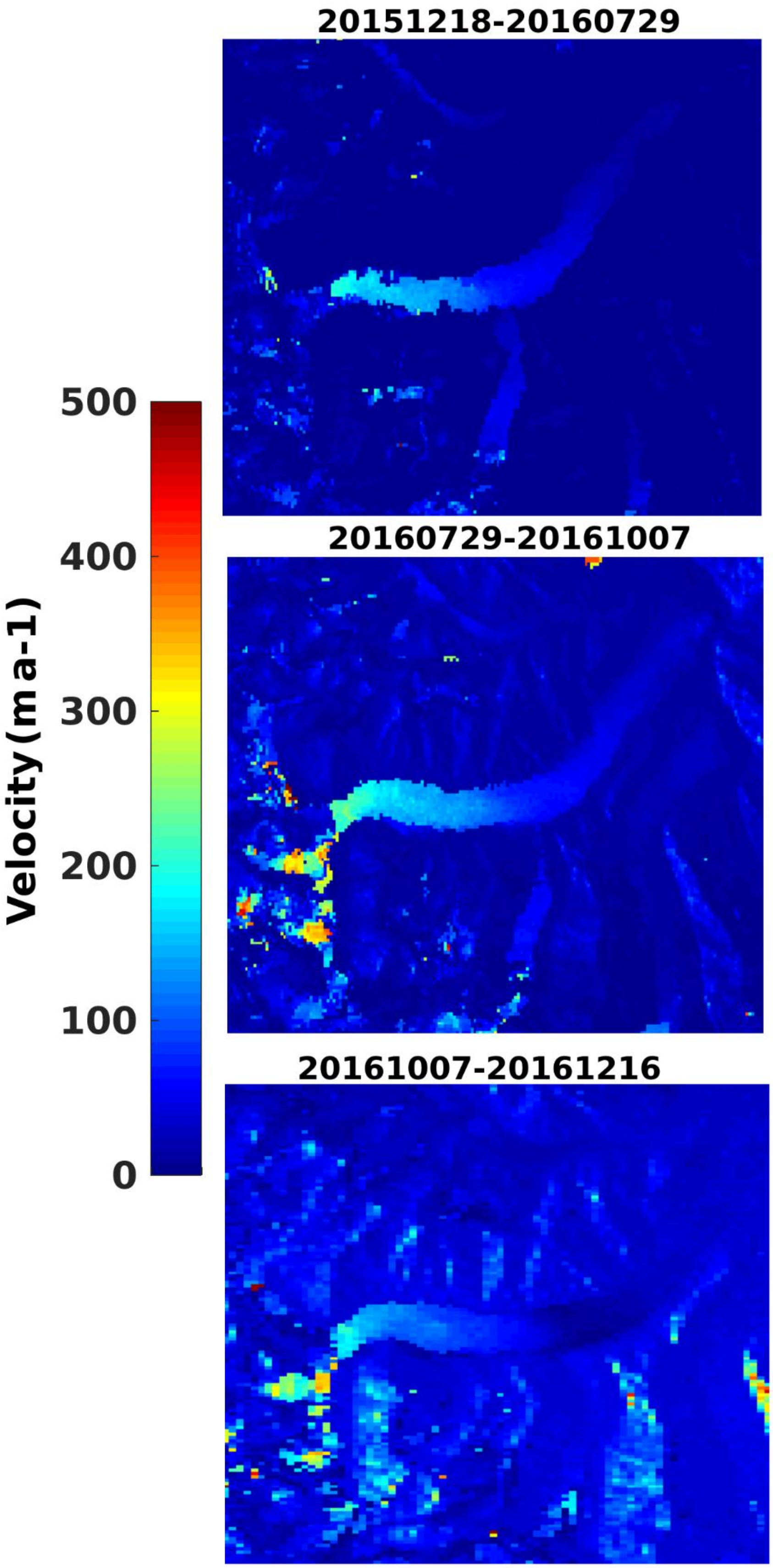

20081211-20090126
20090126-20090613
20090613-20090729
Velocity (m a-1)
500
400
300
200
100
0


20090729-20090913
20090913-20091214
20091214-20100129
Velocity (m a-1)
500
400
300
200
100
0

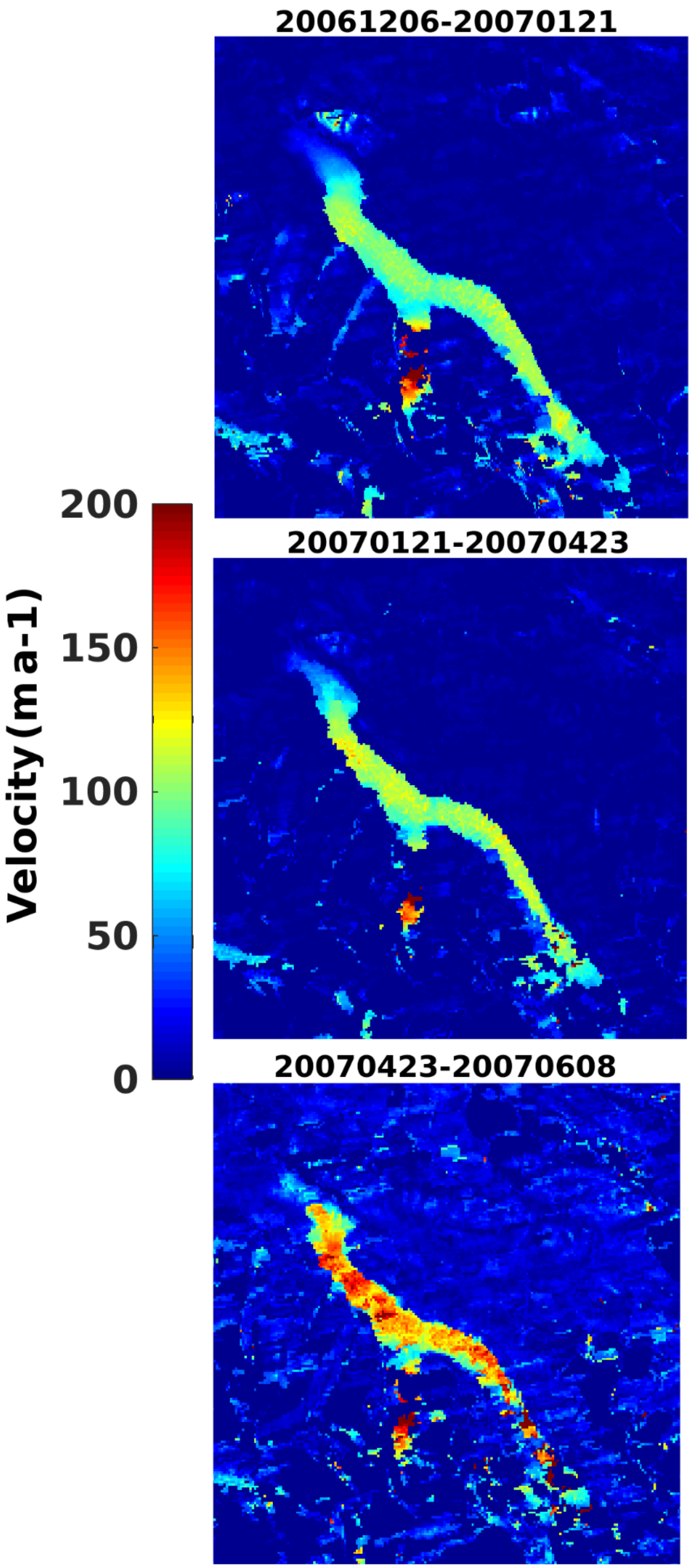

20100129-20100616
20100616-20100801
20100801-20100916
Velocity (m a-1)
500
400
300
200
100
0

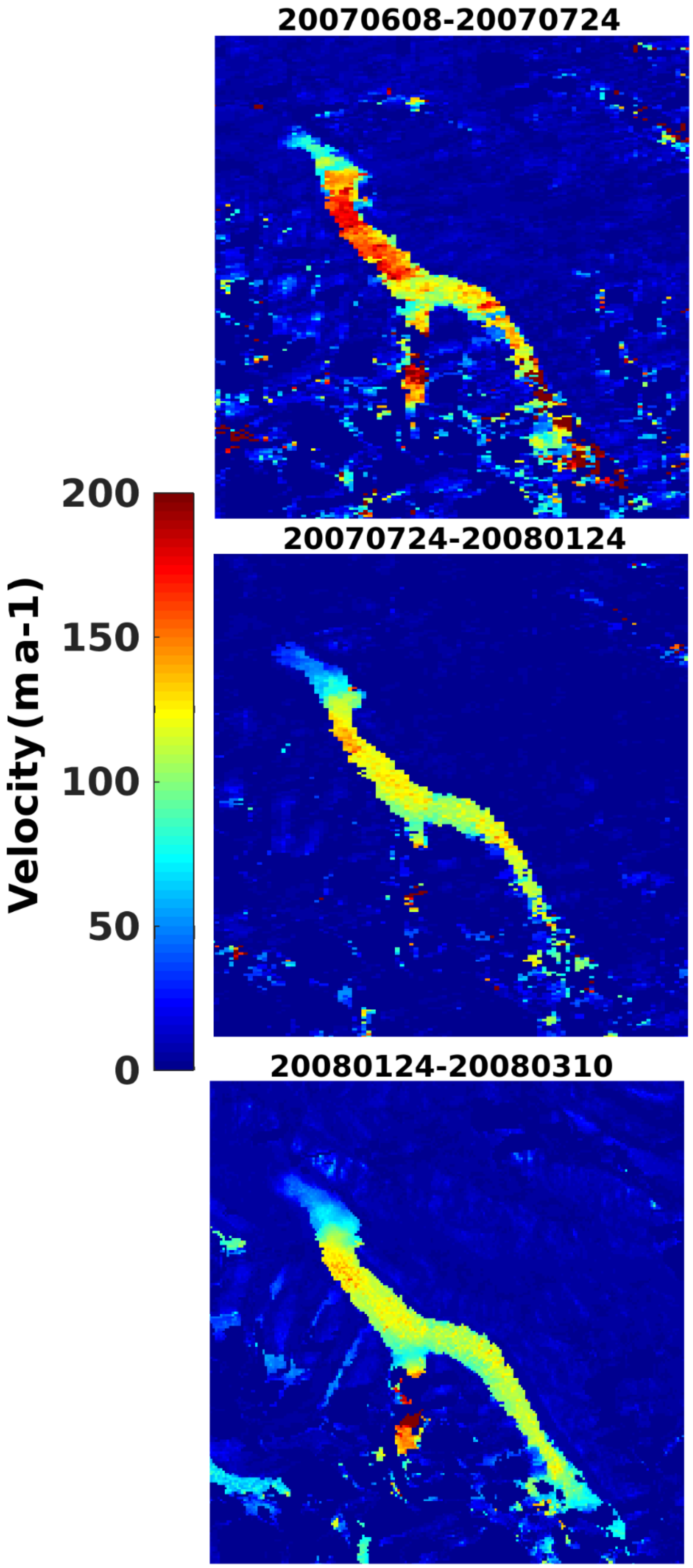

20100916-20101217
20141010-20141219
20141219-20150227
Velocity (m a-1)
500
400
300
200
100
0

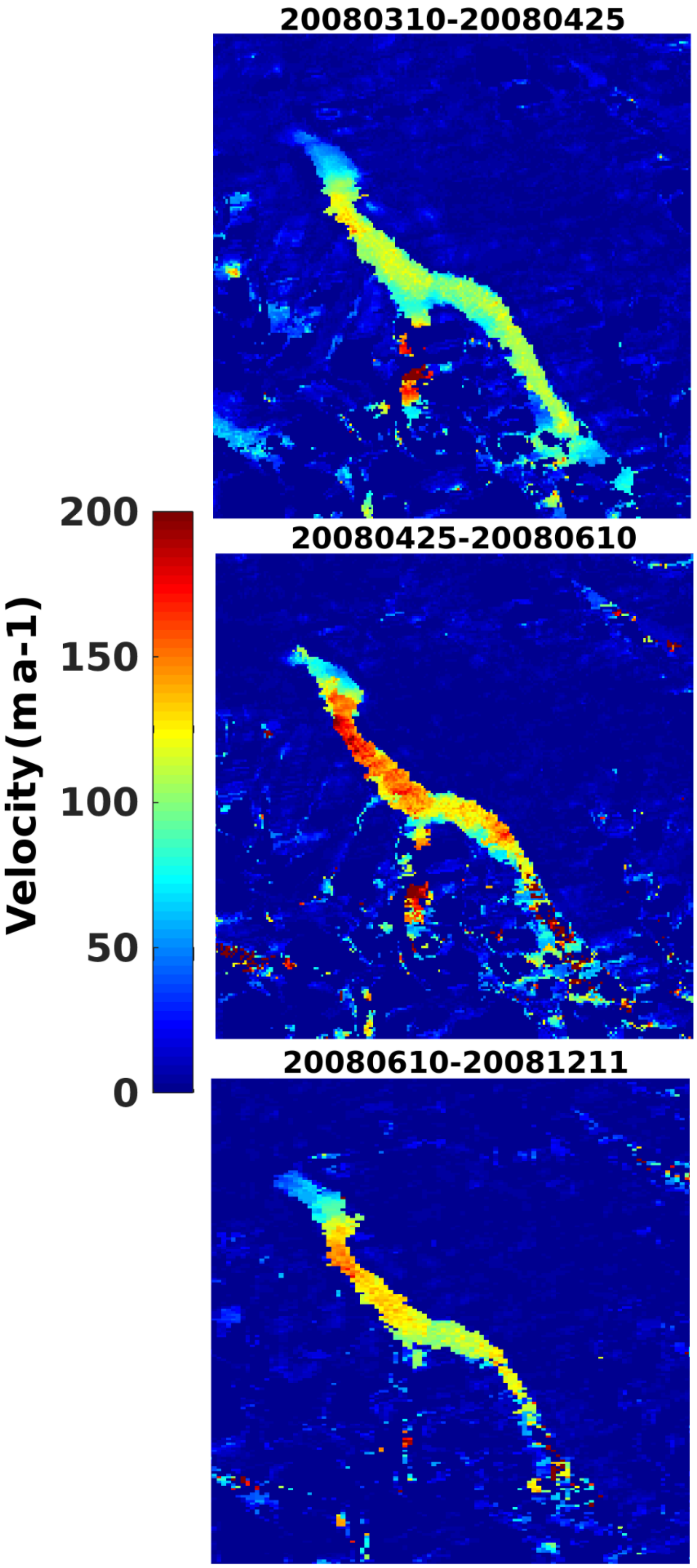

20150227-20150731
20150731-20151009
20151009-20151218
Velocity (m a-1)
500
400
300
200
100
0


20151218-20160729
20160729-20161007
20161007-20161216
500
400
300
200
100
0
Velocity (m a-1)

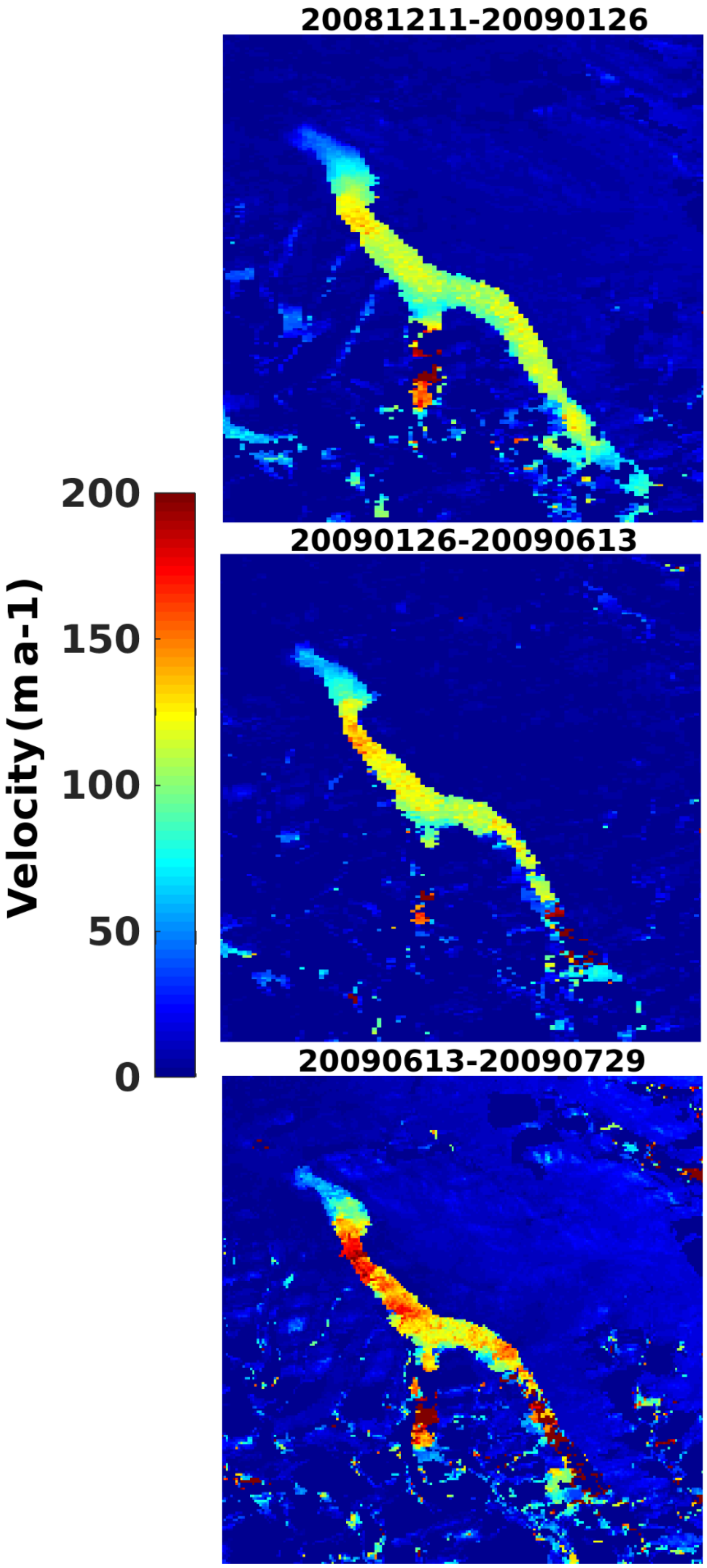
Surface velocity field of the Singkhu Glacier during the study period (Figures S5)

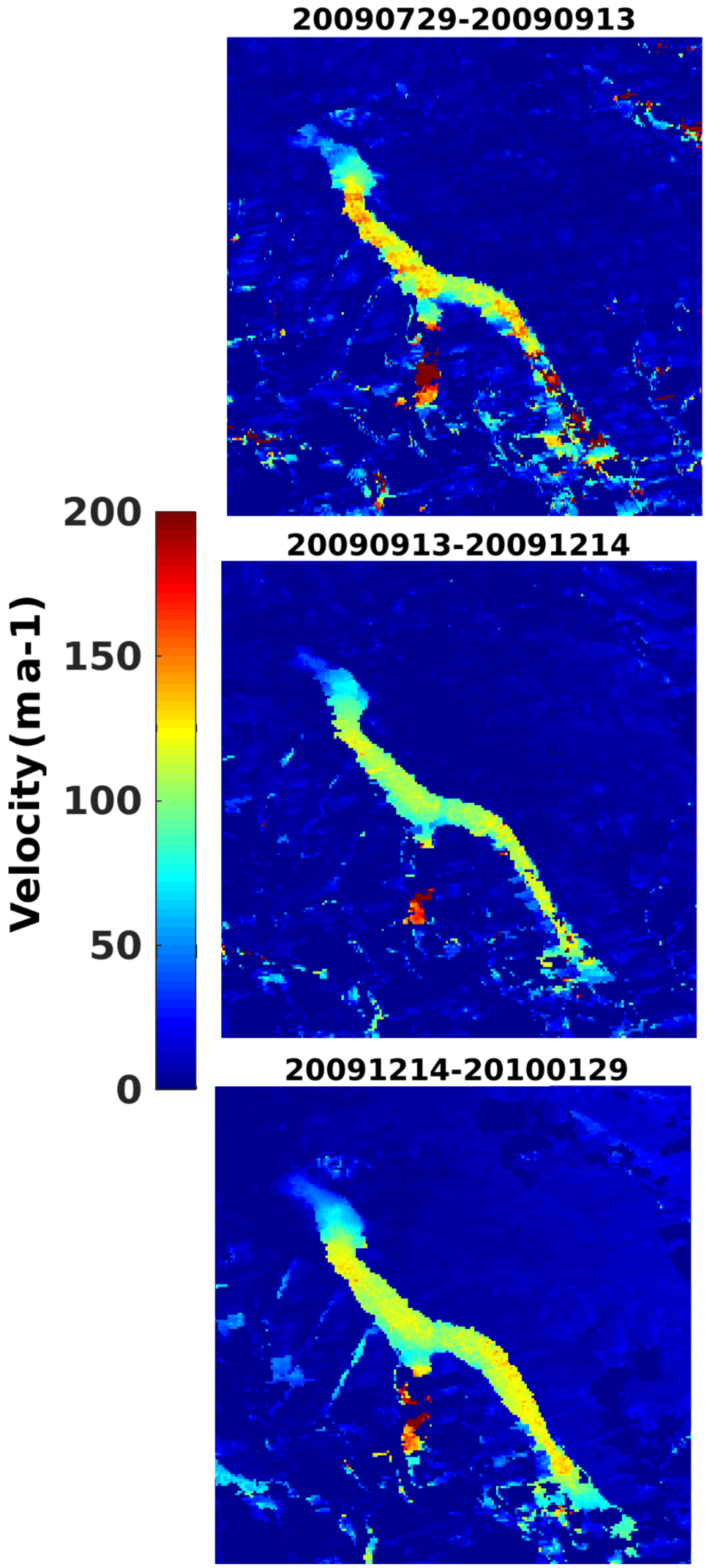

20061206-20070121
20070121-20070423
20070423-20070608
Velocity (m a-1)
200
150
100
50
0

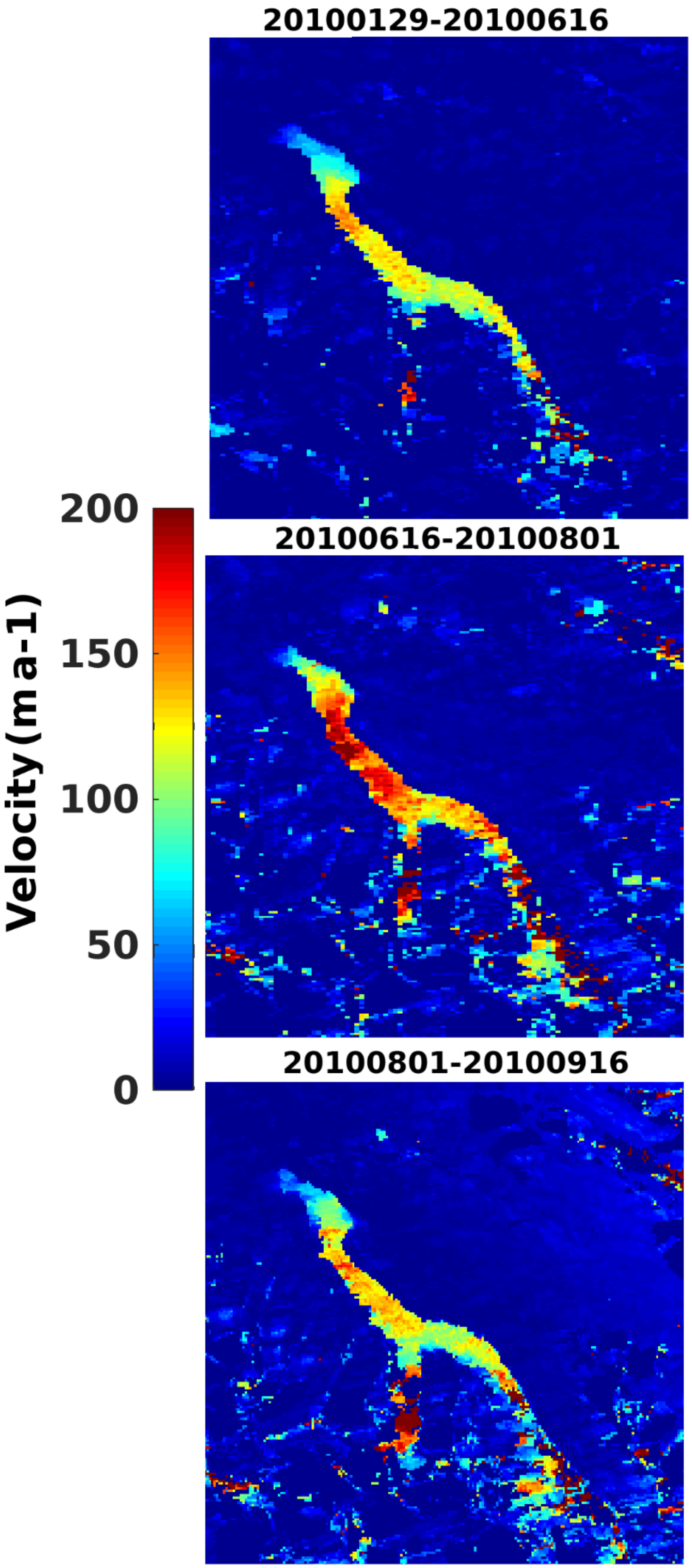

20070608-20070724
20070724-20080124
20080124-20080310
Velocity (m a-1)
200
150
100
50
0

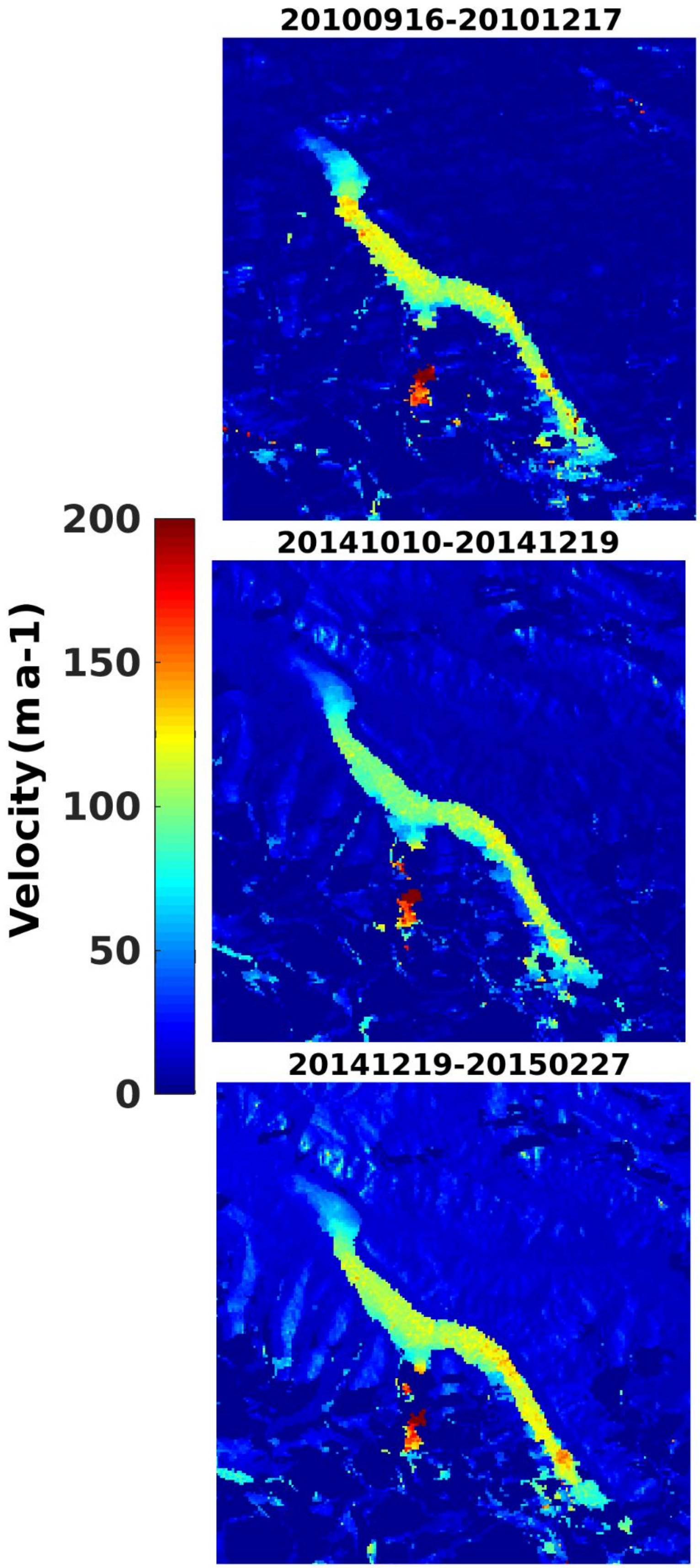


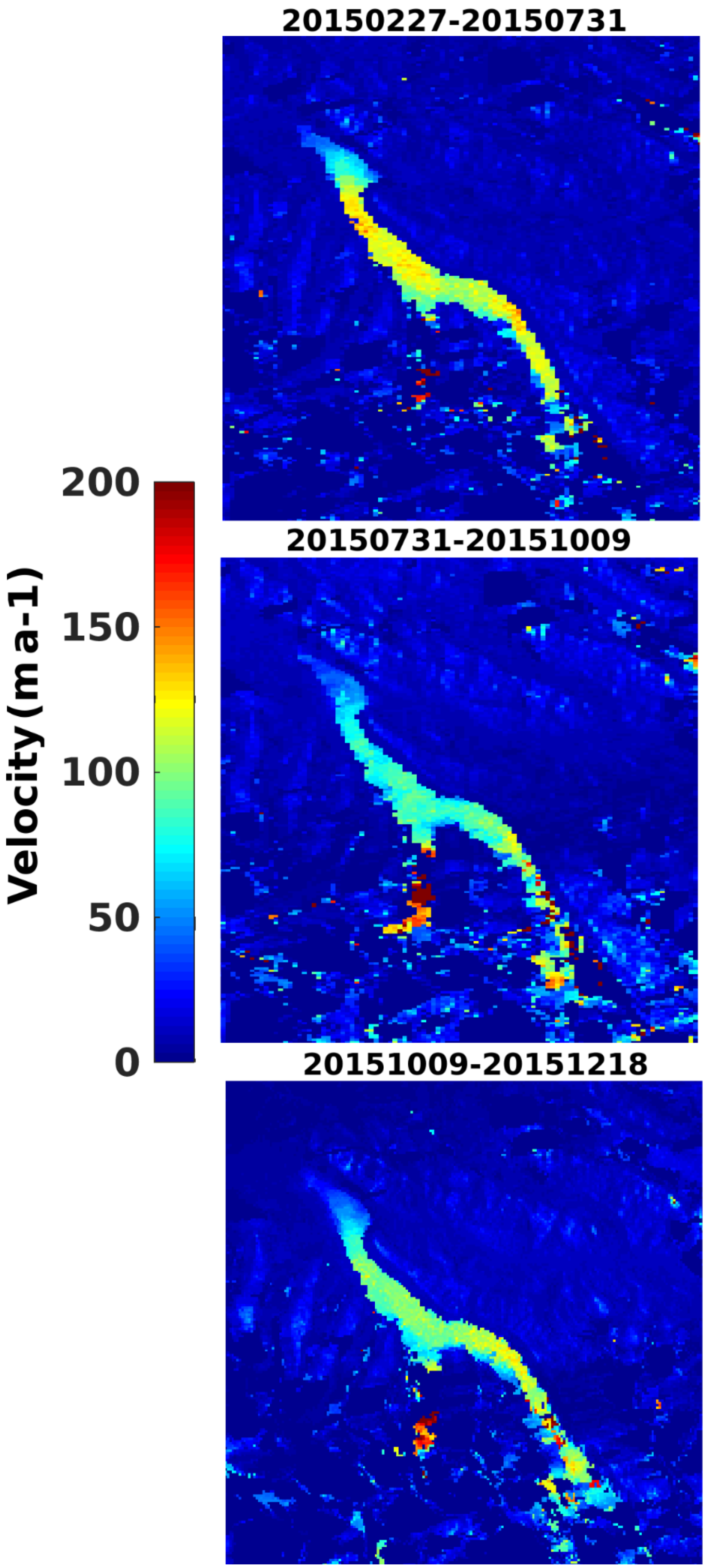
20080310-20080425
20080425-20080610
20080610-20081211
Velocity (m a-1)
200
150
100
50
0

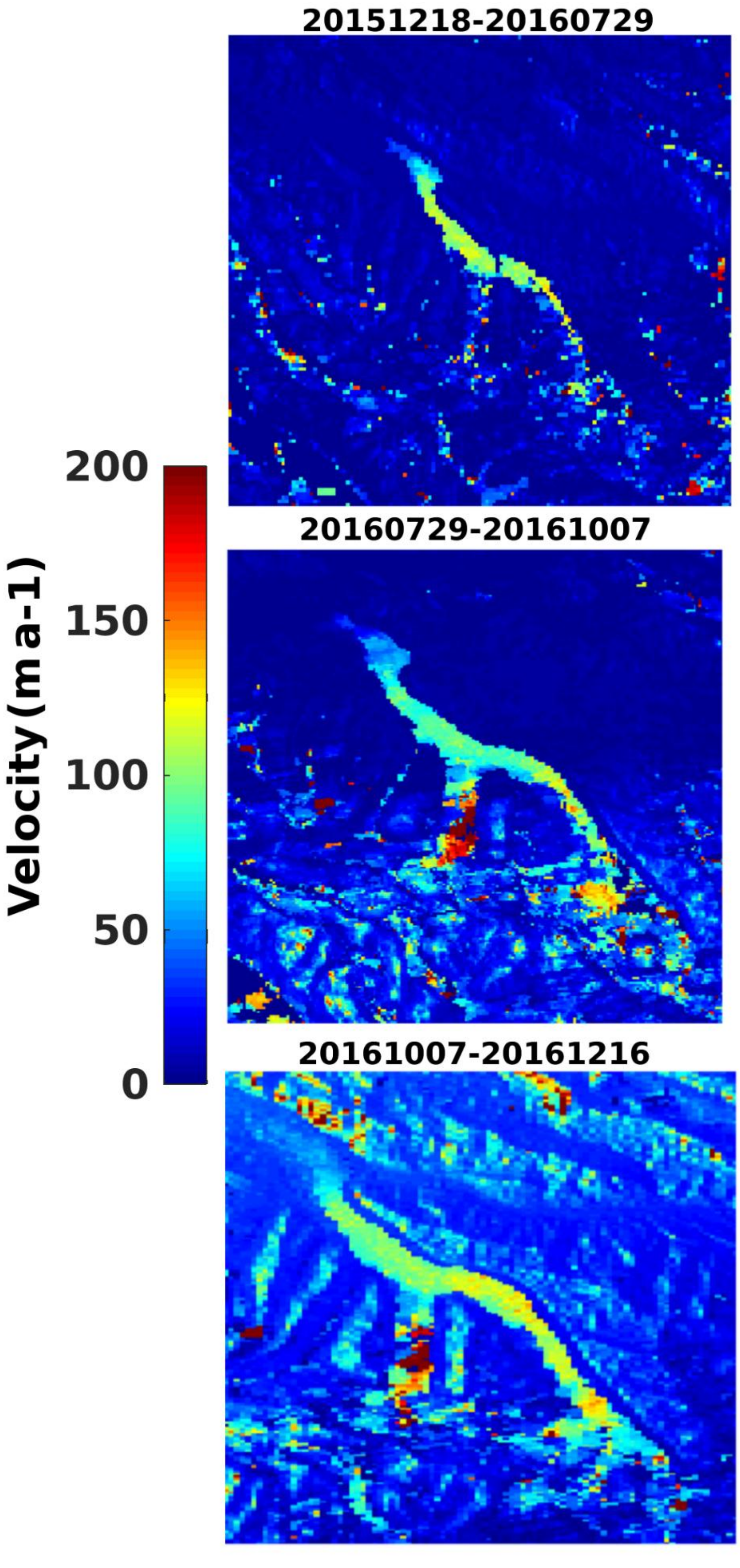

20081211-20090126
20090126-20090613
20090613-20090729
Velocity(m a-1)
200
150
100
50
0

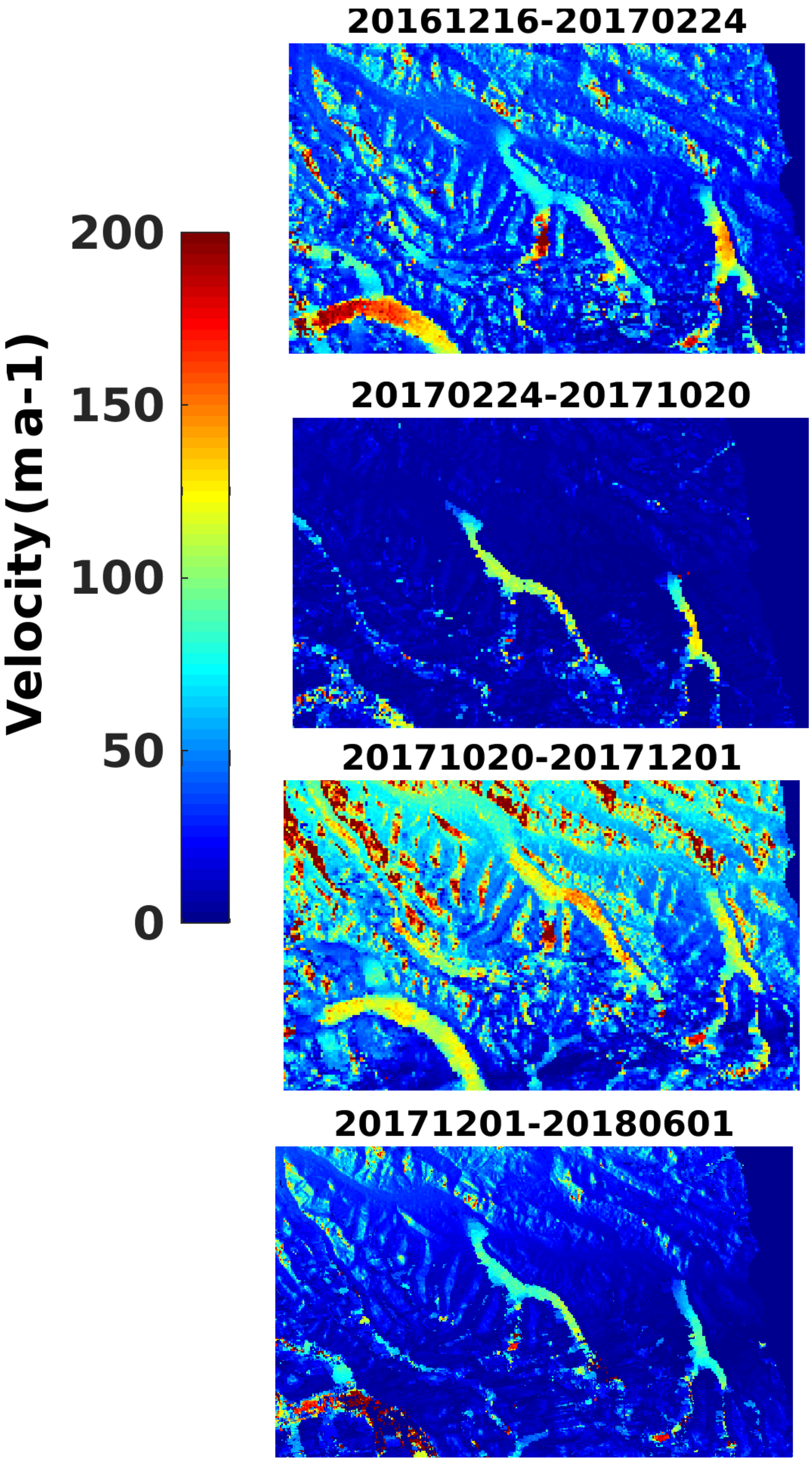

20090729-20090913
20090913-20091214
20091214-20100129
Velocity (ma-1)
200
150
100
50
0


20100129-20100616
20100616-20100801
20100801-20100916
200
150
100
50
0
Velocity (m a-1)


20100916-20101217
20141010-20141219
20141219-20150227
Velocity (m a-1)
200
150
100
50
0


20150227-20150731
20150731-20151009
20151009-20151218
Velocity (m a-1)
200
150
100
50
0


20151218-20160729
20160729-20161007
20161007-20161216
Velocity (m a-1)
200
150
100
50
0


20161216-20170224
20170224-20171020
20171020-20171201
20171201-20180601
Velocity(m a-1)
200
150
100
50
0

References

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