AGU PUBLICATIONS

JGR

Journal of Geophysical Research: Earth Surface

RESEARCH ARTICLE
10.1002/2015JF003515

Special Section:
Glacier Surging and Ice Streaming

Key Points:

- Karakoram glacier surges are heterogeneous in their character
- Karakoram surges do not conform to classic thermal and hydrological surge models
- Controls on surging may differ on an individual glacier basis

Supporting Information:

- Table S1

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Citation:
Quincey, D. J., N. F. Glasser, S. J. Cook, and A. Luckman (2015), Heterogeneity in Karakoram glacier surges, *J. Geophys. Res. Earth Surf.*, *120*, 1288–1300, doi:10.1002/2015JF003515.

Received 3 MAR 2015
Accepted 17 JUN 2015
Accepted article online 18 JUN 2015
Published online 14 JUL 2015

Heterogeneity in Karakoram glacier surges

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Abstract Many Karakoram glaciers periodically undergo surges during which large volumes of ice and debris are rapidly transported downglacier, usually at a rate of 1–2 orders of magnitude greater than during quiescence. Here we identify eight recent surges in the region and map their surface velocities using cross-correlation feature tracking on optical satellite imagery. In total, we present 44 surface velocity data sets, which show that Karakoram surges are generally short-lived, lasting between 3 and 5 years in most cases, and have rapid buildup and relaxation phases, often lasting less than a year. Peak velocities of up to 2 km a^{-1} are reached during summer months, and the surges tend to diminish during winter months. Otherwise, they do not follow a clearly identifiable pattern. In two of the surges, the peak velocity travels down-ice through time as a wave, which we interpret as a surge front. Three other surges are characterized by high velocities that occur simultaneously across the entire glacier surface, and acceleration and deceleration are close to monotonic. There is also no consistent seasonal control on surge initiation or termination. We suggest that the differing styles of surge can be partly accounted for by individual glacier configurations and that while some characteristics of Karakoram surges are akin to thermally controlled surges elsewhere (e.g., Svalbard), the dominant surge mechanism remains unclear. We thus propose that these surges represent a spectrum of flow instabilities and the processes controlling their evolution may vary on a glacier by glacier basis.

1. Introduction

Glacier surges are reported from Canada, Russia, Svalbard, Iceland, Greenland, Alaska, and parts of the Himalaya. Surge-type glaciers undergo cyclical nonsteady flow consisting of two distinct phases [*Meier and Post*, 1969]. The active phase, typically lasting a few months to a few years, is a period of activity during which glacier velocity increases by at least an order of magnitude. The quiescent phase, typically lasting tens to a few hundreds of years, is a period of relative stagnation during which the lower portion of the glacier (known as the receiving area) thins, and mass builds up in an upper, reservoir area. During surges, mass is rapidly transferred from the reservoir to the receiving area, and an advance of the glacier terminus often, but not always, takes place.

Two "types" of glacier surge have long been referred to in the literature, which describe the trigger mechanisms by which an active phase is initiated. In the first, changes in basal temperature promote increased sediment deformation and porosity and a positive feedback between pore water pressure, deformation, and basal flow ensues [*Clarke et al.*, 1984; *Murray et al.*, 2000]. These thermally regulated surges are characterized by an initiation phase that lasts several years before the peak of the surge is reached and a termination phase that comprises several years of deceleration following the peak of the surge. These surges have been observed to begin their acceleration/deceleration independent of any seasonal control. They are mostly recognized in Svalbard [*Murray et al.*, 2003] and the Yukon [*Clarke et al.*, 1984]. In the second, changes in the efficiency of the hydrological system, and thus pore water pressure, trigger the flow instability [*Kamb et al.*, 1985; *Björnsson*, 1998]. Such hydrologically regulated surges are characterized by rapid acceleration and deceleration (i.e., days to weeks long) and tend to initiate during winter months when drainage efficiency is low, terminating during summer months, when drainage efficiency is high. Such events are mostly recognized in Alaska [*Burgess et al.*, 2012; *Lingle and Fatland*, 2003].

Remotely sensed data have provided the foundation for many contemporary studies of surge-type behavior [e.g., *Fatland and Lingle*, 1998; *Murray et al.*, 2003; *Quincey et al.*, 2011; *Mayer et al.*, 2011; *Turrin et al.*, 2013]. Velocity data are derived using cross-correlation feature tracking of either optical imagery or synthetic

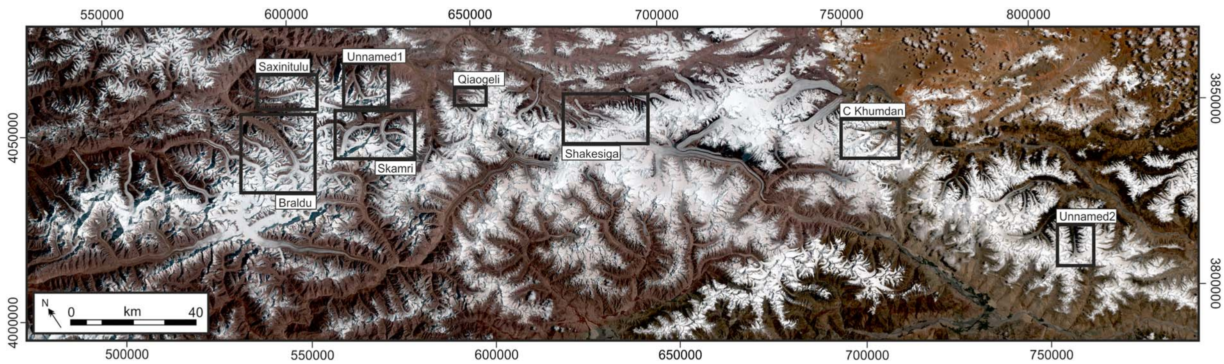

Figure 1. The Karakoram region and the location of the eight glaciers analyzed in this study. Landsat background imagery ©USGS, 2009 + 2010. Coordinates are given in universal transverse Mercator World Geodetic System 84 Zone 43°N. Note that the image has been rotated counterclockwise from true north.

aperture radar imagery, or both, or using interferometry where the surge is slow enough to maintain coherence. Studies have focused on rates of kinematic wave propagation [*Turrin et al.*, 2013], surge return periods [*Quincey and Luckman*, 2014], and the contribution of surging glaciers to tidewater ice fluxes [*Burgess et al.*, 2013]. Many studies have focused on identifying trigger mechanisms [e.g., *Murray et al.*, 2003], but for some regions of the world the mechanics of glacier surging remain poorly understood. This is particularly true in remote terrain, where surges may go entirely undetected or only be recognized once underway. One such region is the Karakoram, Pakistan, which is home to one of the largest concentrations of surge-type glaciers anywhere in the world [*Copland et al.*, 2011] but remains inaccessible for many researchers because of ongoing political tension.

Better quantification of glacier surge dynamics, in particular rates of acceleration and deceleration and how these vary within and between regions, is important to realize if the basal processes that yield such rapid changes are to be understood. In high-elevation regions such as the Karakoram, this also has important implications for landscape evolution, because surging glaciers are effective geomorphic agents [*Humphrey and Raymond*, 1994], and for local water supplies and hazard development, because active and advancing termini can redefine the routing of meltwater, inundate land, and lead to ice-dammed lake development [*Haemmig et al.*, 2014].

In this paper we examine surface velocity data measured on eight Karakoram glaciers during recent surge events (Figure 1 and Table 1). The glaciers vary in character from those with long, debris-covered tongues, the longest of which is the Skamri Glacier at 40 km in length, to those that are short and debris-free (Figure 2). Six of the glaciers are already known to be surge-type: Braldu, Chong Khumdan, Qiaogeli, Skamri, Saxinitulu, and Unnamed1 [*Copland et al.*, 2011; *Gardelle et al.*, 2012; *Rankl et al.*, 2014]. The other two glaciers in our data set are not known to have surged in the past. The aim of the paper is to assess whether surging glaciers of the Karakoram show dynamic characteristics that conform with long-standing dynamic models of surge-type behavior. We use multiple sources of remote sensing data to determine

Table 1. Selected Characteristics of Glaciers in This Study[a]

Glacier Name	Latitude (dec deg)	Longitude (dec deg)	Maximum Elevation (meter above sea level (asl))	Minimum Elevation (m asl)	Length (km)	Debris Covered	Aspect (deg)	Last Known Surge	Reference (if Applicable)
Braldu	36.143	75.865	6300	3970	34	yes	0	unknown	*Copland et al.* [2011]
Chong Khumdan	35.183	77.679	6370	4720	20	yes	110	1927–1928	*Copland et al.* [2011]
Qiaogeli	35.967	76.456	7067	4777	9.5	partly	310	1990–2000	*Copland et al.* [2011]
Saxinitulu	36.281	75.943	6286	4600	16.5	no	290	unknown	*Gardelle et al.* [2012]
Shakesiga	35.715	76.851	7030	4420	26	no	320	unknown	-
Unnamed1	36.178	76.202	6956	4340	14	partly	10	unknown	*Rankl et al.* [2014]
Unnamed2	34.605	77.978	6435	4746	11	partly	20	unknown	-
Skamri	36.055	76.178	6700	3989	40.5	partly	90	1978?	*Copland et al.* [2009]

[a]Elevations and lengths are the approximate values, and aspect is taken to be the dominant flow direction.

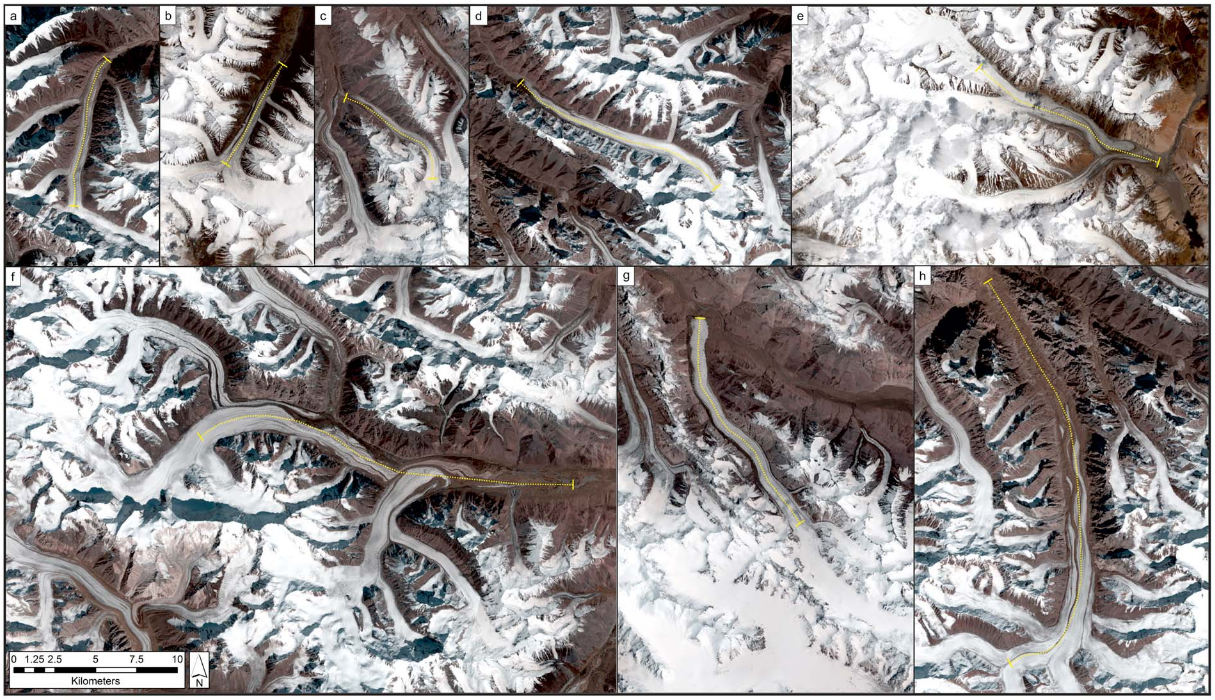

Figure 2. Detailed view of the eight glaciers and the centerline profiles used to extract velocity data (shown in Figure 4): (a) Unnamed1, (b) Unnamed2, (c) Qiaogeli, (d) Saxinitulu, (e) Chong Khumdan, (f) Skamri, (g) Shakesiga, and (h) Braldu. In each case the profile is taken from the maximum terminus position reached during each glacier surge and from the terminus moving upglacier.

Karakoram surge dynamics across an unprecedented number of events and present a glaciological interpretation of the likely controlling mechanisms.

2. Study Area: Karakoram Glaciers

The glaciers of the Karakoram have been intensively studied in recent years because, in contrast to most glacierized regions of the world, they have been gaining mass [*Kääb et al*., 2012; *Gardelle et al*., 2012]. The majority of the glaciers have either stable termini positions, or have been advancing [*Scherler et al*., 2011; *Bhambri et al*., 2013], partly as a consequence of increased regional precipitation [*Janes and Bush*, 2012] and partly because of glacier surging [*Paul*, 2015]. Previous work has suggested a preponderance of surge-type behavior in glaciers between 12 and 25 km in length [*Hewitt*, 1969] and those fed by tributary glaciers [*Hewitt*, 2007], although several recent studies have shown that many of the smaller glaciers in the region also surge [*Rankl et al*., 2014; *Paul*, 2015]. Surges can lead to kilometer-scale advances of glacier termini over very short time scales, i.e., months to years.

Previous work focusing on the triggers of Karakoram surges have arrived at conflicting conclusions [*Quincey et al*., 2011; *Mayer et al*., 2011]. On one hand, Karakoram glacier surges have been suggested to be thermally rather than hydrologically controlled, coinciding with high-altitude warming from long-term precipitation and accumulation patterns [*Quincey et al*., 2011; *Quincey and Luckman*, 2014]. On the other, observations and modeling from a single surge event invoked a change in hydrological conditions as the main trigger mechanism [*Mayer et al*., 2011]. There is some evidence that glacier surges are increasing in frequency in the region [*Copland et al*., 2011], although recent data have suggested otherwise [*Rankl et al*., 2014]. Furthermore, despite satellite observations now being available for more than three decades, return periods are still poorly constrained. Estimates and observations normally cite typical return periods on the order of 25–40 years [*Guo et al*., 2013; *Copland et al*., 2011], although historical observations of the Khurdopin Glacier suggest a slightly shorter return period of ~20 years [*Mason*, 1930; *Quincey and Luckman*, 2014].

3. Methods

Multitemporal velocity fields were calculated by cross-correlation feature tracking [*Luckman et al*., 2007]. This method has been repeatedly shown to produce high-quality results on Himalayan and Karakoram glaciers because of the abundance of surface features associated with debris-cover and surge-type flow [*Quincey et al*., 2009, 2011; *Mayer et al*., 2011]. Satellite images were sourced from Landsat TM (Thematic Mapper), Landsat ETM+ (Enhanced Thematic Mapper), Landsat OLI (Operational Land Imager), ALOS Advanced Visible and Near-Infrared Radiometer (AVNIR), and Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) sensors (Table S1 in the supporting information) to give as dense a data set as possible through each of the surges. The feature-tracking approach has been well-described elsewhere so we provide a summary of our approach here. In the case of the AVNIR and ASTER data, the first step was to orthorectify the images using the automated function within ENVI 5.1, which is based on sensor model and digital elevation data. All Landsat imagery was provided at L3, with the orthorectification already carried out by U.S. Geological Survey (USGS). The images were then coregistered on an individual glacier scale to correct for remaining misalignment. We used coarse windows of 128×128 (pattern size) and 256×256 (search area) to achieve this. Horizontal ground displacements were extracted using a Fourier-based correlation technique [*Luckman et al*., 2007] with search windows of between 24×24 to 64×64 pixels (pattern size) and 32×32 to 128×128 pixels (search area).

Errors in the resulting displacement data arise from misregistration of the two satellite images and the precision of the algorithm used. Our coregistration is subpixel and is therefore likely to be similar to the ~5 m accuracy quoted by *Lee et al.* [2004] when considering multitemporal Landsat 7 ETM+ images acquired on the same path and row. The correlation technique is affected by changes in crevasses and surface debris patterns through time and space as well as the potential for mismatches of surface features. To mitigate against the latter errors, resultant displacement data were filtered using signal-to-noise ratio >7.0 as the primary indicator of the quality of the match. We also removed extreme values (i.e., above a stipulated maximum threshold in each data set) and removed matches that deviated from the dominant glacier flow direction by $>30°$. This left only the most robust patch correlations, for which the measurements themselves are expected to be of subpixel accuracy.

To provide an indication of the uncertainty (σ) in the remaining velocity values we used the following equation, modified from *McNabb et al.* [2012]:

$$\sigma = 365 \frac{\left(C_{\text{pix}} + C_{\text{match}}\right)\Delta x}{\Delta t}$$

where C_{pix} is the uncertainty in coregistration in pixels (p), C_{match} is the uncertainty in the matching algorithm in pixels (p), Δx is the image resolution in meters, and Δt is the time interval between the image pair in days. Using typical values of 0.5 p for C_{pix} and C_{match} we estimate uncertainties that in the majority of cases are $<100\ \text{m a}^{-1}$ (Table S1). It should be noted that even when uncertainties $>100\ \text{m a}^{-1}$ these data coincide with the peak surge velocities, and the measured displacements still far exceed the potential errors.

To aid interpretation of the surge dynamics, surface debris structures were mapped for every glacier using time-separated optical satellite images in ArcGIS. Features mapped include glacier extent, areas of surface debris, and associated surface debris structure.

4. Results

Fourty-four velocity fields were derived through the eight glacier surges (Figure 3). It should be noted that our derived velocity data are generally restricted to the ablation area, so our analysis does not focus on dynamics in the accumulation zone. Centerline profiles show the magnitude and timing of each event as it impacts the lower part of the glacier (Figure 4; note that we do not plot error bars here to avoid obscuring data patterns). The maximum velocity recorded in any of the data sets was ~2 km a^{-1}, while the slowest surge reached just 300 m a^{-1}; in all cases the peak surge velocities exceeded those in the buildup period by at least 1 order of magnitude and in some cases, 2 orders of magnitude.

While it is difficult to identify exactly when each of the surges initiated, some insight can be drawn from looking at the differences between individual profiles. In the case of the first unnamed glacier (Unnamed1; Figure 4e), the lowermost 7 km of the glacier was flowing at $<400\ \text{m a}^{-1}$ during the summer months of

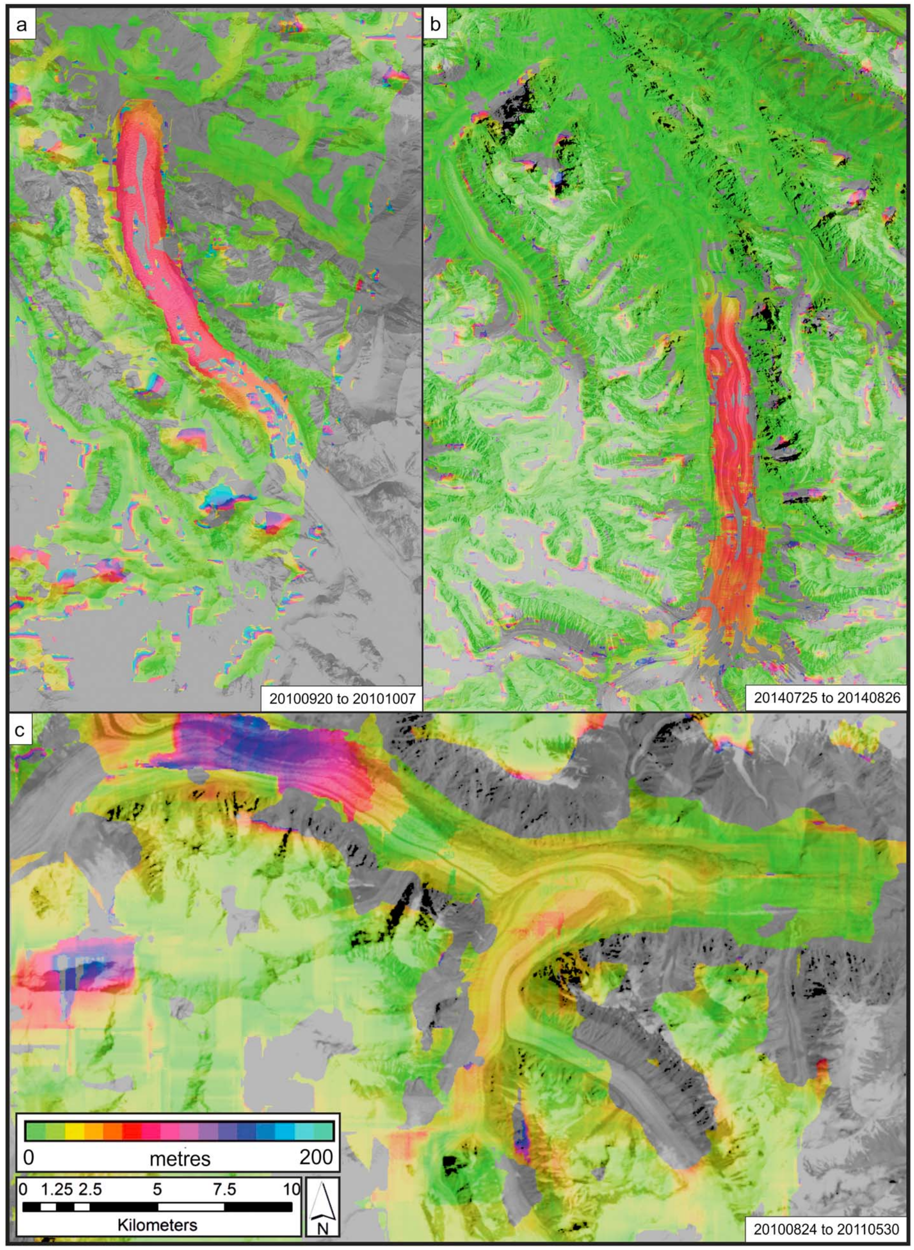

Figure 3. Selected filtered velocity fields for (a) Shakesiga, (b) Braldu, and (c) Skamri.

2009, but had accelerated to $>1500\,\mathrm{m\,a^{-1}}$ by the early summer months of 2010, indicating that sometime during the winter months of 2009, the switch between slow and fast flow took place. Similarly, the Shakesiga surge was in its infancy during the late summer of 2009 with the glacier flowing at $<400\,\mathrm{m\,a^{-1}}$ (Figure 4f), but had reached its maximum velocity of ~2000 $\mathrm{m\,a^{-1}}$ by midsummer of 2010, again indicating that the switch took place during winter months. In the case of the second unnamed glacier (Unnamed2; Figure 4g), the surge appears to have been developing during the summer months of 2006 with a zone of fast flow between 5 and 7 km from the terminus and actually receded during the following winter months before switching again to fast flow in the summer of 2007 and reaching its maximum velocity of $800\,\mathrm{m\,a^{-1}}$ during this period. The initiation phase is missing in the available data for several of the other surges, but the data from the Skamri Glacier (Figure 4h) also suggest that the switch to fast flow took place more toward the summer season than the winter. In all cases it appears that the initiation phase was months to years long.

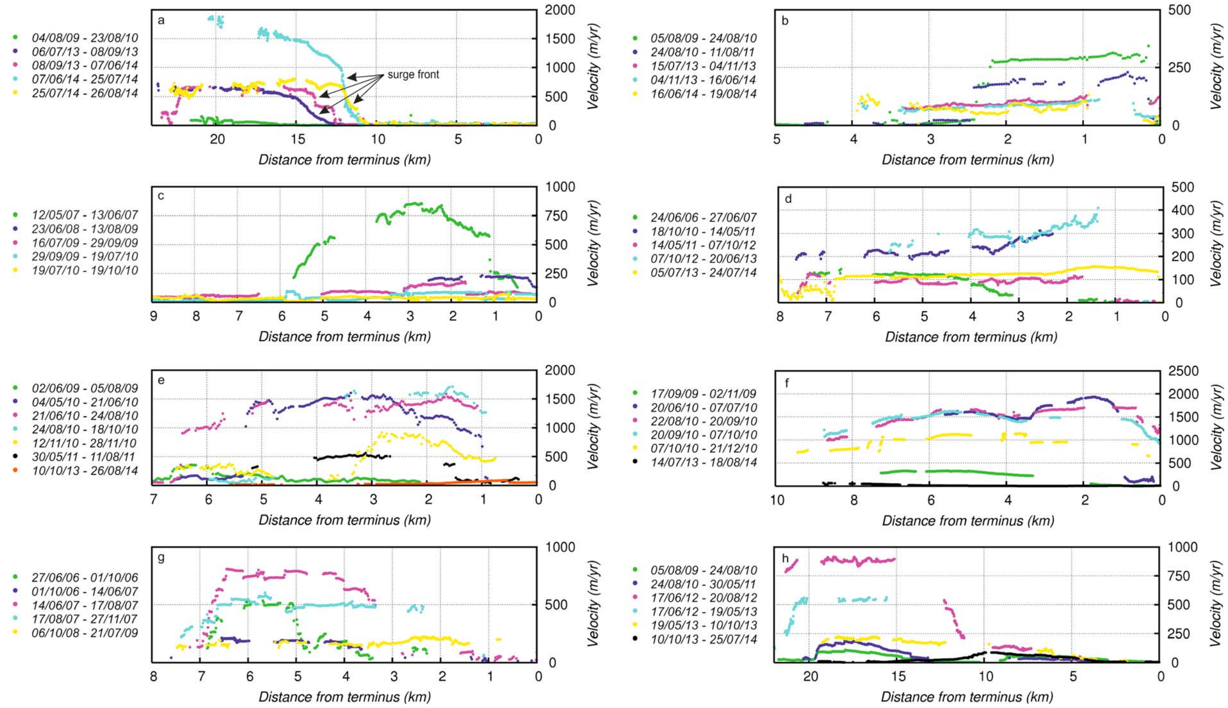

Figure 4. Centerline velocity profiles characterizing the dynamic evolution of surges on each of the eight glaciers in the study: (a) Braldu, (b) Qiaogeli, (c) Chong Khumdan, (d) Saxinitulu, (e) Unnamed1, (f) Shakesiga, (g) Unnamed2, and (h) Skamri. For error estimation, see Table S1. The axes scales are not directly comparable. Note that surge velocities are between 1 and 2 orders of magnitude greater than quiescent velocities in each case and the clear downglacier migration of a surge front in the Braldu data set (labeled).

The termination periods also appear to have been variable in their timing. Perhaps the best defined is that of Unnamed1 (Figure 4e), where the surge was clearly active during the summer months of 2010 but began decelerating at the start of the following winter, i.e., in the November 2010 data set. The Shakesiga surge follows a similar dynamic (Figure 4f), with the glacier decelerating to 1100 m a^{-1} during the early winter of 2010 having peaked at almost twice this velocity in the immediately preceding summer months. In several other cases the termination phase was slow to develop, and thus, identifying when the switch from fast to slow flow took place becomes difficult. Nevertheless, it appears that the termination phase was longer than the initiation phase in the data sets where observations for both are possible (four of the eight data sets—Shakesiga, Unnamed1, Unnamed2, and Skamri; Figures 4e–4h). In all four of these cases, the total surge lasted for between 3 and 5 years; in the fifth (Saxinitulu; Figure 4d), the surge is still ongoing, 8 years after initiation.

In common with previous observations on the Kunyang Glacier [*Quincey et al*., 2011], at least two of the currently studied glacier surges are characterized by a downglacier propagation of the velocity peak. We interpret this to represent the surge front, although we have no surface elevation data to confirm its topographic expression. In the case of the Braldu surge, there is a clear velocity wave that propagates downglacier at approximately 2 km a^{-1} at the height of the surge (Figure 4a). There is a less-clear front in the Unnamed1 data set, but during the summer of 2010, the peak velocity did migrate downglacier and its arrival at the glacier terminus coincided with a deceleration both around the terminus and upglacier. There are also hints of a surge front in both the Chong Khumdan and Skamri data sets but based only on limited data. In contrast, other glaciers show a very different dynamic, with the surge affecting almost the whole glacier coincidentally. The Shakesiga data set shows this most clearly (Figure 4f), with a uniform increase in flow of between 1000 and 2000 m a^{-1} across the entire glacier length. A similar, but less pronounced, increase is also visible in the Qiaogeli and Saxinitulu surges (Figures 4b and 4d). The Unnamed1 data set shows characteristics of both surge styles, with a generally monotonic acceleration/deceleration affecting the lowermost ~7 km of ice, but also showing some evidence of a surge front.

Figure 5. Before and during the surge of Saxinitulu Glacier: (a) 5 August 2009, (b) 14 May 2011, and (c) 26 August 2014. The surge peaked in 2013, but the glacier terminus is still advancing in early 2015 imagery.

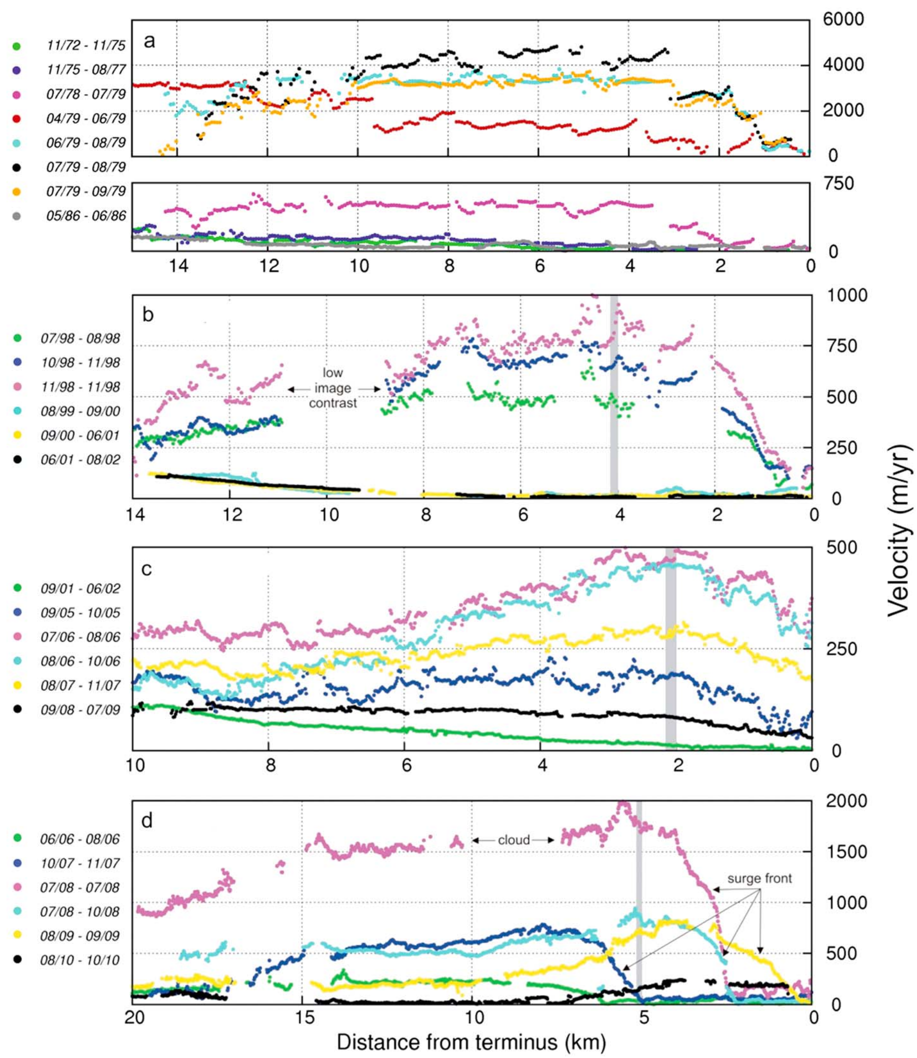

Figure 6. Velocity data for four previously published surges on (a) Khurdopin Glacier (during the late 1970s [*Quincey and Luckman*, 2014]), (b) Khurdopin Glacier (during the late 1990s), (c) Gasherbrum Glacier, and (d) Kunyang Glacier [*Quincey et al.*, 2011].

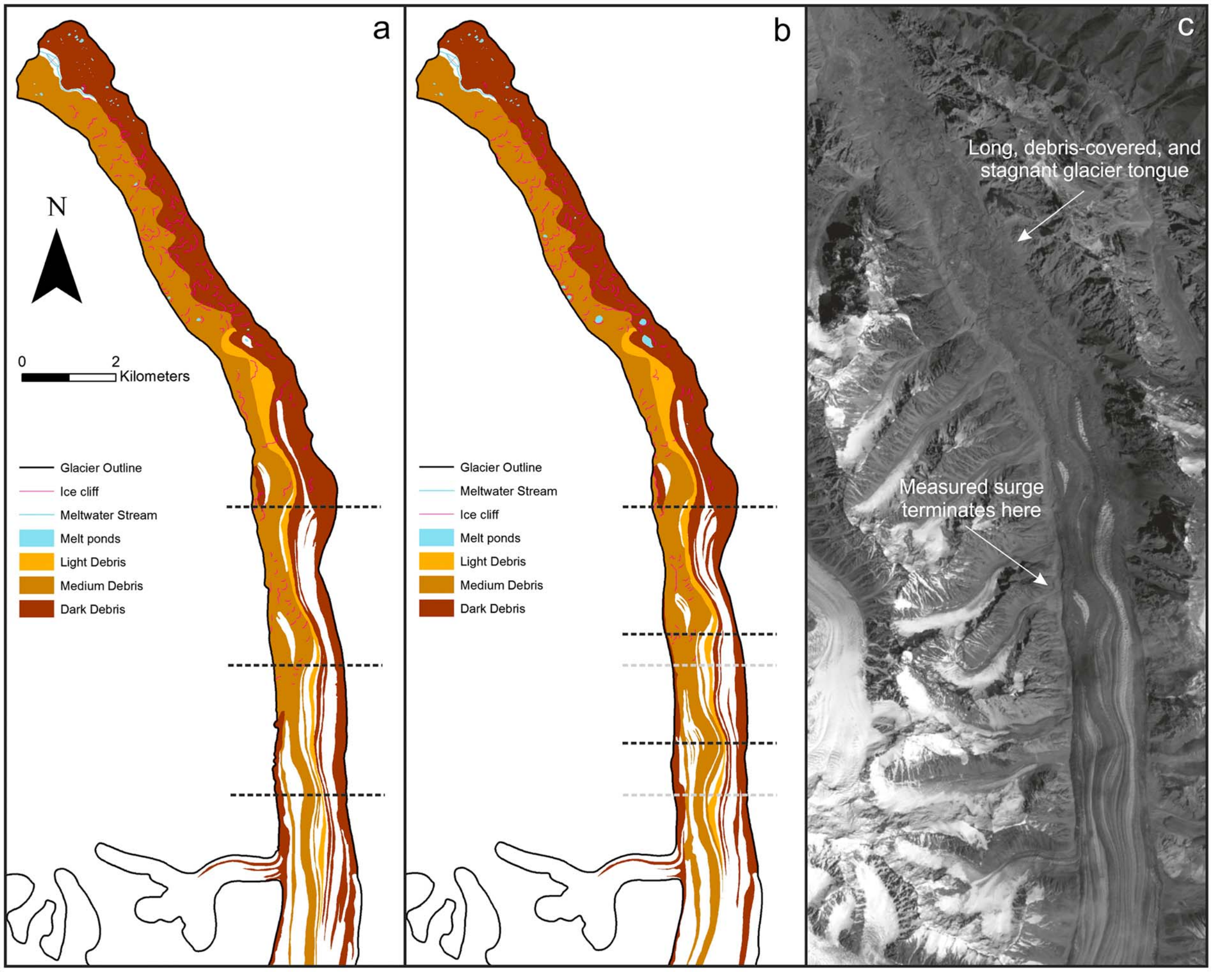

Figure 7. The geomorphic context of the Braldu surge: (a) 5 July 2013, (b) 25 August 2014, and (c) Landsat ETM+ panchromatic imagery acquired 25 August 2015 for reference. The black dashed lines indicate the prominent surface features and their relative positions in each data set. The grey dashed lines in the August 2014 data set denote the relative position of each feature in the July 2013 data set. Note the long debris-covered tongue that provides a major obstacle to the downglacier propagation of the surge front.

Several of the shorter glaciers experienced frontal advances of up to 2 km (Figure 5), whereas surges within the longer glaciers were mostly confined to the existing glacier area. The Braldu surge, although still technically ongoing, does not look likely to impact the lowermost 10 km of debris-covered ice. Similarly, the Skamri surge looks to have terminated approximately 10 km from the terminus. The Shakesiga surge resulted in a small frontal advance of several hundred meters but not sufficient to override the main valley river and abut the opposing valley wall. Both of the unnamed glaciers as well as the Saxinitulu Glacier and the Qiaogeli Glacier advanced by several kilometers during their surges; indeed, the Saxinitulu Glacier is still advancing at ~100 m/yr having already advanced almost 2 km from its original terminus position.

5. Discussion

Previous studies focusing on Karakoram surges have suggested that both thermal and hydrological controls may be responsible for their initiation [*Quincey et al*., 2011; *Mayer et al*., 2011]. Evidence that has supported the thermal switch hypothesis includes the apparently random timing of the initiation phase and its length, which usually lasts several years, as opposed to the <0.5 years observed in other regions [*Kamb et al*., 1985], as well as a surge front identified in one data set (Kunyang Glacier [*Quincey et al*., 2011]) that may have represented the boundary between the thawed and frozen bed [cf. *Fowler et al*., 2001]. Numerical modeling has been used to explain the propagation of a similar surge front on the Gasherbrum

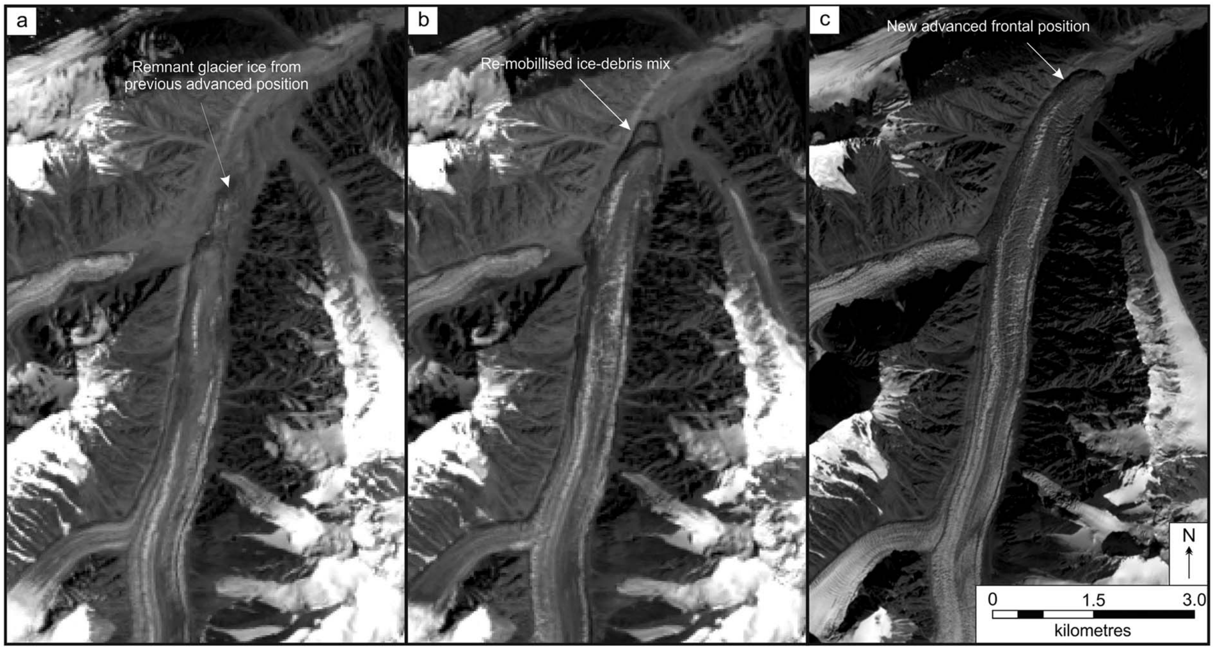

Figure 8. Evolution of the Unnamed1 surge: (a) 5 August 2009, (b) 24 August 2010, and (c) 10 October 2013. Note the former glacier position approximately 1 km downvalley of the active terminus in 2009, and the way in which that ice-debris mix is overridden by the most recent surge event. Note also the advancing terminus to the true left of the main glacier tongue, which stops just short of becoming an active tributary to the main surge.

Glacier using concepts of glacier sliding with cavitation and subglacial hydrological switching and to explain modulation waves (small-amplitude velocity peaks) identified in the feature-tracked velocity data [*Mayer et al*., 2011]. Coupled with these previous observations, multitemporal velocity data now exist for 12 Karakoram surges (Figures 4 and 6), including one duplicate, Khurdopin Glacier [*Quincey et al*., 2011; *Quincey and Luckman*, 2014]. These combined data show that no single dynamic model can describe the surges of the Karakoram; indeed, they represent a broad spectrum of velocity patterns. Some are characterized by a peak-velocity wave propagating downglacier, which we interpret as a surge front; others are characterized by more uniform and simultaneous acceleration over the full glacier length.

Mayer et al. [2011] identified a surge front in their Gasherbrum velocity data, and *Quincey et al.* [2011] reported similar observations on the Kunyang Glacier. Traveling waves have been observed during many previous glacier surges and have been linked to surges controlled by both hydrology [*Kamb et al*., 1985; *Fowler*, 1987] as well as thermal changes [*Fowler et al*., 2001]. In the case of the former, the surge front is thought to represent the transition between an efficient tunnel drainage system promoting flow by deformation downglacier of the front and an inefficient linked-cavity system promoting flow by sliding upglacier of the front. It has been suggested that there may be a seasonal signal to hydrologically controlled surge front propagation [*Turrin et al*., 2013; *Raymond*, 1987], with deceleration during summer months when subglacial channelization reduces water pressure and acceleration during winter conditions when the basal hydrology is inefficient. In the case of the thermal switch theory, the boundary is thought to be between warm-ice upglacier of the front and cold-ice downglacier of the front. According to *Clarke* [1976], the cold ice is immobile and frozen to its bed during quiescence. The critical element in terms of whether a surge initiates appears to be the thickness and permeability of the underlying sediment layer [*Fowler et al*., 2001], and where there is no restriction to flow at the margin, the surge front may be entirely absent.

The Braldu surge is relatively short-lived, and given the temporal resolution of the observations, it is difficult to determine any seasonal signal in the propagation of its surge front (Figure 4a). However, the fact that its downglacier progression is inhibited by immobile, and probably cold, ice is clear to see in both the velocity data and in the geomorphological interpretation, which illustrates a long, stagnant, debris-covered tongue (Figure 7). The other data set in which a surge front may be present is Unnamed1. This glacier is particularly

Table 2. Surge Characteristics for All 12 Events in the Karakoram That Have Been Observed With Multitemporal Velocity Data[a]

Source	Glacier	Surge Front	Terminus Advance	Winter Initiation	Summer Termination	Monotonic Acceleration	Initiation Shorter Than Termination	Peak Velocity in Summer
This study	Braldu	••••	No presence	No data	••••	••	No data	••••
	Chong Khumdan	No presence	No presence	No data	No data	No data	No data	•
	West Qogori (Qiaogeli)	No presence	••••	No data	•	•••	No data	No data
	Saxinitulu	No presence	••••	No data	No data	•••	No data	No data
	Shakesiga	No presence	••	••••	No presence	••••	•	••••
	Unnamed1	•••	••••	••••	No presence	•••	••	••••
	Unnamed2	No presence	••••	No presence	••••	•	•	••••
	Skamri	No presence	No presence	No presence	•	••	•	••••
Quincey and Luckman [2014]	Khurdopin (1970s)	No presence	No presence	••	••••	•••	No data	••••
Quincey et al. [2011]	Khurdopin (1990s)	No presence	No presence	No data	No data	•••	••	••
	North Gasherbrum	No presence	No presence	•	••	••••	••••	••••
	Kunyang	••••	••••	••	••••	No presence	••••	••••

[a]The presence of each characteristic is denoted by • = weak presence to •••• = strong presence, where there is insufficient data to assess the characteristic we state "no data."

interesting because the surge appears to have overridden debris or dead ice that is a remnant of a previous advanced glacier position (Figure 8). In both cases, therefore, significant obstacles impeded the surge. The same is true for the Kunyang surge identified in *Quincey et al.* [2011]; the Kunyang Glacier showed extensive areas of thermokarst presurge indicating stagnant or slow-moving ice, and the main glacier into which the Kunyang feeds, the Hispar Glacier, is known to be slow-flowing [*Rankl et al.*, 2014] and thus provides a further obstacle to fast-flowing ice. It is therefore possible that these surge fronts could simply be a consequence of the individual glacier configurations rather than representing a thermal or drainage boundary as has been invoked elsewhere [*Fowler et al.*, 2001; *Kamb et al.*, 1985].

In contrast, several of the gathered data sets show a much more uniform and spatially coincident acceleration, akin to that observed at Monacobreen in Svalbard [*Murray et al.*, 2003]. The equivalent end-member in our Karakoram data appears to be the Shakesiga data set, although the Saxinitulu and Qiaogeli surges and previous profiles for the Khurdopin Glacier and the Gasherbrum Glacier (Figure 6) are similarly characterized. In such cases, the lack of a surge front could be accounted for by a thermal activation front propagating faster than ice flow, and consequently, no buildup of fast-flowing ice is apparent [*Fowler et al.*, 2001]. Similarly, the dynamic evolution of surges observed on smaller glaciers in our data set (Unnamed1 and Unnamed2) also conform to theoretical analysis of thermal triggers in that the greatest acceleration is observed as the glacier front begins to advance. It is possible that in these latter cases, the thermal activation wave has already reached the terminus by this point, and as the glacier forefield is warm, the ice can advance and accelerate unabated [cf. *Fowler et al.*, 2001]. Alternatively, if the hydrological system is uniform across the glacier bed, a coincident and glacier-wide switch from efficient to inefficient drainage could explain the monotonic acceleration [*Björnsson*, 1998].

A collective analysis of the 12 velocity data sets we have now derived for Karakoram Glacier surge events shows no clear pattern to suggest that a single trigger mechanism operates in the region (Table 2). Some of the observed characteristics conflict directly with those reported on hydrologically controlled surges elsewhere, indicating that perhaps thermal regulation may be dominant: (1) the shape of the buildup, active surge, and termination phases of the Karakoram surges contrast with those reported from Alaskan glaciers [e.g., *Burgess et al.*, 2012], where hydrology appears to be a dominant control. In Alaskan Glacier surges, the termination phase is much more abrupt than the initiation phase, tending to last several days (or even hours) as opposed to months (or even years) [*Kamb et al.*, 1985]. In the Karakoram, on many glaciers the termination phase can last for years (Figure 9), suggesting that in these cases the mechanisms operating are different to

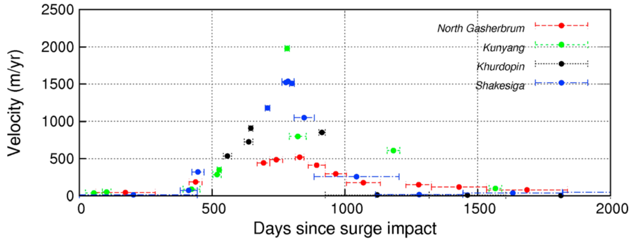

Figure 9. Surge evolution of previously measured events in the Karakoram [*Quincey et al.*, 2011] and the Shakesiga event measured here. Note that the shape of the acceleration and deceleration resembles those with a thermal control in Svalbard [*Murray et al.*, 2003] but that the relatively short overall surge period (~600–900 days in each case) is more akin to the sudden acceleration and deceleration of hydrologically controlled surges in Alaska [*Kamb et al.*, 1985].

those operating in Alaska. (2) The length of the buildup phase can be of the order of several years in the case of Karakoram surges as opposed to several months as would be predicted by the hydrological surge initiation model. Indeed, *Mayer et al.* [2011] cited this as the main conflict between their observed and modeled dynamics, suggesting that the 3 year buildup phase of the Gasherbrum surge greatly exceeded the expected time to switch between an efficient and inefficient drainage system. (3) The timing of the initiation and termination phases appears to be independent of any seasonal control. Hydrologically controlled surges tend to initiate during winter months and terminate during summer months; the Karakoram surge data presented here and elsewhere do not conform to this pattern. (4) Peak velocities are consistently reached during summer months in Karakoram surges. If the surge control was hydrological, we might expect there to be a deceleration during summer months [cf. *Kamb et al.*, 1985] when the basal hydrology would be relatively efficient. (5) There is no evidence of subglacial water either at the margins or within crevasses on the surging glaciers of the Karakoram, which would support a theory of elevated water pressure being a major control on surging [e.g., *Jiskoot et al.*, 2001]. (6) There have been no observations of short-lived, large-scale velocity variations that were a feature of the Variegated Glacier surge and other hydrologically controlled surges [e.g., *Kamb et al.*, 1985].

Intriguingly, however, two main features of the observed Karakoram surges do not conform to thermally controlled events elsewhere: (1) the return periods of Karakoram glacier surges are notably shorter than those reported for thermally controlled surges elsewhere, being of the order of several decades rather than several centuries [*Quincey and Luckman*, 2014]. In all eight cases studied here, the last known surge was pre-1992 (confirmed by the satellite record), so we can report that their return periods are at least 15 years. (2) Karakoram surges tend to last for much shorter periods than those in Svalbard, for example (~3–5 years, as opposed to ~10 years). In extreme cases, they can last as little as 1–2 years, as with the Shakesiga Glacier (Figure 9). This short-lived switch from slow to fast flow resembles Alaskan-type surges more than the Svalbard-type.

The dynamics of Karakoram Glacier surges do not therefore fit neatly into the well-cited dynamic classification of thermal and hydrologically controlled surges. There are many remaining unknowns in the Karakoram region that are all likely to play a role in surge magnitude and frequency and may help to explain the inconsistency. The greatest gap in Karakoram Glacier knowledge relates to glacier basal conditions, in terms of their thermal characteristics, their composition, and their roughness. Previous work has suggested that cold ice may predominate at high elevations and around the margins of the larger debris-covered glaciers [e.g., *Quincey et al.*, 2009] but based only on seasonal variations in surface velocity. Indeed, given the extreme relief of the Karakoram Mountains and the elevation range over which glaciers can be found, it is likely that many different thermal regimes are present, making conventional classes such as warm, cold, and polythermal, devised for other contexts, inappropriate for these glaciers [*Hewitt*, 2014]. Similarly, little is known about whether the beds of these surge-type glaciers are hard or soft, although field observations have identified thicknesses of several meters of basal debris [*Owen and Derbyshire*, 1989], indicating that soft sediment may well underlie at least some of the glaciers in the region but not necessarily all. Even less is known about their roughness, which may determine the rate of sliding and mass flux if the underlying sediment is immobile [*Zoet and Iverson*, 2015]. Finally, the region is geologically complex, with most surge-type glaciers crossing two or more major formations [*Hewitt*, 1998], and possibly underlain by spatially variable geothermal heat flow [*Chamberlain et al.*, 1995].

Karakoram glaciers are situated at much higher elevation than those in other surge-prone regions of the world and are generally shorter and much steeper [*Hewitt*, 1998]. It might be reasonably expected that the overall surge cycle may be much more frequently occurring and shorter lived simply because the accumulation areas of the Karakoram glaciers cannot store vast volumes of ice as can their polar (e.g., Svalbard) counterparts. Based on the evidence presented here, we suggest that the thermal, sedimentological, and geomorphological characteristics of Karakoram glaciers may vary even on a glacier by glacier basis, and thus, the classic thermal and hydrological classification is not appropriate in the Karakoram context. We propose that Karakoram Glacier surges have individual dynamic characteristics and cannot be collectively characterized. The implication of this is that the controlling processes are variable and likely reflect a spectrum of flow instabilities.

6. Conclusions

Using cross-correlation feature tracking applied to optical satellite imagery we have made a significant addition to existing data describing the temporal and spatial evolution of Karakoram glacier surges. These data demonstrate that (1) Karakoram surges are generally short-lived, lasting between 3 and 5 years from initiation to termination, although longer in some cases. (2) The initiation and termination phases are rapid, lasting months to years, and do not appear to be seasonally controlled. (3) The frontal advances of some small surging glaciers can exceed 2 km over several years of surging. (4) Surge fronts are present in some Karakoram surges but may simply reflect individual glacier configurations. (5) Uniform acceleration and deceleration across the whole glacier surface more typically characterizes these fast-flow events. (6) Maximum velocities are on the order of 2 km a^{-1} as has been reported in previous work. (7) Surging tends to peak, and often decelerate, during summer months. The dynamic evolution of Karakoram surges does not therefore fit neatly within either of the classically cited thermal or hydrological models of surging, suggesting that no single process is responsible for their instability. Their heterogeneity adds weight to the idea that glacier surge events are part of a continuum bound at one end by normal, slow flow, and at the other by permanent, fast flow, with a combination of hydrological and thermal basal processes determining their dynamic evolution.

Acknowledgments
The authors would like to acknowledge the provision of ALOS AVNIR imagery through research agreement 1008 (Quincey) and ASTER imagery under the NASA-affiliated researcher program. All data used in the formulation of this manuscript are available from the first author on request.

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