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 down-glacier, usually at a rate of one to two 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 datasets, which show that Karakoram surges are generally short-lived (between 3 and 5 years in most cases), have rapid build-up and relaxation phases (often less than a year), and generally reach peak velocity during summer 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. Others are characterised by high velocities that occur simultaneously across the entire glacier surface and acceleration and deceleration is 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 accounted for by individual glacier geometries, 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

classic thermal and hydrological surge classifications are not appropriate in the Karakoram context because the surges cannot be collectively categorised. The implication of this is that regional triggers may also not be singularly defined, and may even differ on a glacier by glacier basis.

1. Introduction

Glacier surges are reported from the Canadian and Russian High Arctic, Svalbard, Iceland, Greenland, Alaska and parts of the Himalaya. These surge-type glaciers undergo cyclical non-steady 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 (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 (thermally-regulated and hydrologically controlled) 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 characterised by several years of acceleration before the peak of the surge is reached (often termed the initiation phase), several years of deceleration following the peak of the surge (often termed the termination phase), and tend to begin their acceleration/deceleration independent of any seasonal control. They are mostly recognised 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 characterised by rapid acceleration and deceleration (i.e. days to weeks long), and tend to initiate during winter months (a time of drainage inefficiency) and terminate during summer months (when efficiency is increased). Such events are mostly recognised in Alaska (Burgess et al., 2012; Lingle and Fatland, 2003).

Remotely sensed data have provided the foundation for many contemporary studies of surge-type behaviour (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 either cross-correlation feature tracking (of either optical imagery or synthetic aperture radar imagery, or both) or interferometry (where the surge is slow enough to maintain coherence), and studies have focussed 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., 2014). Many studies have focussed 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 recognised once underway. One such region is the Karakoram, Pakistan, which is home to one of the largest concentrations of surging 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 (magnitude of, and spatial variability in acceleration and deceleration) and how they differ within and between regions are important to realise 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 (in terms of erosion/deposition) as well as local water supplies and hazard development (in terms of land inundation, ice/rock avalanching from surging masses, and ice-dammed lake development). Thus, the aim of this paper is to augment the limited surge data we have for the region already (Quincey et al., 2011; Mayer et al., 2011) with measured changes in surface velocity on eight further Karakoram valley glaciers during recent surge events. These new results indicate there are distinct similarities between Karakoram surges and those documented in Svalbard (e.g. Murray et al., 2003), but that some dynamic characteristics are more consistent with a hydrological control, as has been suggested elsewhere (Mayer et al., 2011). We conclude that Karakoram surges do not fit neatly within long-standing dynamic models of surge-type behaviour, and suggest that unstable flow should be viewed as a continuum rather than as a binary classification.

2. Study area: Karakoram glaciers

The majority of glaciers in the Himalaya are receding and have lost significant mass since at least 1970, despite thick debris cover (Bolch et al., 2011; Kääb et al., 2014). Glacier wastage is spatially heterogeneous, and is linked to both topography and climate (Fujita and Nuimura, 2011). More than 65% of the monsoon-influenced glaciers studied by Scherler et al. (2011) were observed to be receding. However, heavily debris-covered Karakoram glaciers with stagnant low-gradient terminus regions typically have stable fronts and indeed some glaciers are advancing as increased surface debris cover retards glacier melt (Scherler et al., 2011). Other Karakoram glaciers advance periodically during surges, when velocities increase rapidly to rates between one and two orders of magnitude greater than during quiescence (Hewitt, 1969). Previous work has suggested a preponderance of surge-type behaviour in glaciers between 12 and 25 km in length (Hewitt, 1969) and those fed by tributary glaciers (Hewitt, 2007). The season of Karakoram glacier surge initiation varies, and surges have been shown to develop gradually over several years (Quincey et al., 2011). These can lead to km-scale advances of glacier termini over very short (monthly to annual) timescales.

Previous work focussing 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 et al., 2014). On the other, observations and modelling from a single surge event invoked a change in hydrological conditions as the main trigger mechanism (Mayer et al., 2011). There is some consensus that glacier surges are increasing in frequency in the region, but return periods are poorly constrained. Estimates and observations normally cite typical return periods of 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).

Here we present data on glacier velocity and changes in the surface character of eight Karakoram glaciers through recent surges (Figure 1; Table 1). The glaciers vary in character from long, debris-covered tongues, the longest of which is the Skamri Glacier (~40 km) located directly to the east of the Shimshal Valley, to short (i.e. < 15 km length), debris-free glaciers that are unnamed, at least in the scientific literature (Figure 2). Five of the glaciers are already known to be surge-type (e.g. Braldu, Chong Khumdan, West Qogori, Skamri and Saxinitulu; Copland et al., 2011; Gardelle et al., 2012), while the other three have not previously been identified as surge-type. The contrasting dynamics of the eight surges combined with their distinct surface geomorphologies provides the opportunity to evaluate the processes controlling surge initiation and development in more detail than has previously been possible.

3. Methods

Multi-temporal velocity fields were calculated by cross-correlation feature-tracking (Strozzi et al., 2002). 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; Mayer et al., 2011; Quincey et al., 2011). Satellite images were sourced from Landsat TM, Landsat ETM+, Landsat OLI, ALOS AVNIR and ASTER sensors (Table 2) to give as dense a dataset 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 (based on sensor model and digital elevation data) within ENVI 5.1. All

Landsat imagery was provided at L3, with the orthorectification already carried out by USGS. The images were then co-registered on an individual glacier scale to correct for remaining misalignment. We used coarse windows of 128 x 128 (pattern size) and 256 x 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 x 24 to 64 x 64 pixels (pattern size), and 32 x 32 to 128 x 128 pixels (search area).

Errors in the resulting displacement data arise from mis-registration of the two satellite images and the precision of the algorithm used. Our co-registration is sub-pixel, and is therefore likely to be similar to the ~5 m accuracy quoted by Lee et al. (2004) when considering multi-temporal 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 mis-matches of surface features. To mitigate against the latter errors, resultant displacement data were filtered using signal-to-noise ratio as the primary indicator of the quality of the match. We also removed extreme values (i.e. above a stipulated max threshold) and removed matches that did not conform to the general flow direction of the glacier, defined manually by the user. This left only the most robust patch correlations, for which the measurements themselves are expected to be of sub-pixel 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):

$$\sigma = 365\frac{C_{pix}C_{match}\Delta x}{\Delta t}$$

where C_{pix} is the uncertainty in co-registration in pixels, C_{match} is the uncertainty in the matching algorithm, Δx is the image resolution in metres, and Δt is the time interval between the image pair in days. The highest uncertainty is thus associated with short (16-day) data separations (Table 2).

However, it should be noted that as these data coincide with the peak surge velocities, 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. Centreline profiles show the magnitude and timing of each event as it impacts the lower part of the glacier (Figure 4; n.b. we do not plot error bars here to avoid obscuring data patterns). The maximum velocity recorded in any of the datasets was ~2 km a^{-1} and in all cases the peak surge velocities exceeded those in the build-up period by at least one, and in some cases two, 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), there was relatively slow flow during the summer months of 2009, but the surge was fully developed 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, but had reached its maximum velocity by mid-summer of 2010, again indicating the switch took place during winter months. In the case of the second unnamed glacier (Unnamed2), the surge appears to have been developing during the summer months of 2006 and actually receded during the following winter months before switching to fast flow in the summer of 2007. The initiation phase is missing in the available data for several of the other surges, but the data from the Skamri Glacier also suggest that the switch to fast flow took place more towards the summer season than the winter. In all cases it appears that the initiation phase was months to years long.

The termination periods also appear to have been variable in their timing. Perhaps the best defined is that of Unnamed1, where the surge was clearly active during the summer months of 2010 but began decelerating at the start of the following winter (in the November dataset). The Shakesiga surge follows a similar dynamic, with the surge appearing to diminish in the early winter of 2010 having peaked 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 datasets where observations for both are possible (four of the eight datasets – Shakesiga, Unnamed1, Unnamed2 and Skamri). In all four of these cases, the total surge lasted for between 3 and 5 years; in a fifth (Saxinitulu) the surge is still ongoing, eight 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 characterised by a down-glacier 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 down-glacier at approximately 2 km a^{-1} at the height of the surge (Figure 4). There is a less-clear front in the Unnamed1 dataset, but during the summer of 2010 the peak velocity did migrate down-glacier and its arrival at the glacier terminus coincided with a deceleration both around the terminus and up-glacier. There are also hints of a surge front in both the Chong Khumdan and Skamri datasets, 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 dataset shows this most clearly, with a uniform increase in flow across the entire glacier length. A similar, but less pronounced, increase is also visible in the Saxinitulu and Qiaogeli surges. The Unnamed1 dataset 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.

Several of the smaller (<15 km) glaciers experienced major frontal advances (Figure 5) whereas surges within the larger (>10 km) 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 metres, 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 kilometres during their surges; indeed the Saxinitulu Glacier is still advancing at ~100 m per year having already advanced 1 km from its original terminus position.

5. Discussion

Previous studies focussing on Karakoram surges have suggested 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 (usually 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 dataset (Kunyang Glacier; Quincey et al., 2014) that may have represented the boundary between the thawed and frozen bed (cf. Fowler et al., 2001). Numerical modelling has been used to explain the propagation of a similar surge front on the Gasherbrum 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, multi-temporal velocity data now exist for twelve Karakoram surges (Figures 4 and 6), including one duplicate, Khurdopin Glacier (Quincey et al., 2011; Quincey and Luckman, 2014)). These combined data suggest that at least two types of surge exist in the Karakoram: the first is characterised by a peak-velocity wave (which we interpret as a surge front) propagating down-glacier; the second is characterised 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. (2014) reported similar observations on the Kunyang Glacier. Travelling waves have been observed during many previous glacier surges, and have been linked to both hydrological trigger (Kamb et al., 1985; Fowler, 1987) as well as thermal trigger mechanisms (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 (down-glacier) and an inefficient linked-cavity system promoting flow by sliding (up-glacier). 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 contrasting (hydrologically inefficient) winter conditions. In the case of the thermal switch theory, the boundary is thought to be between warm-ice (up-glacier) and cold-ice (down-glacier). 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 (or lack of signal) in the propagation of its surge front. However, the fact that its down-glacier progression is inhibited by immobile (and probably cold) ice is clear to see in both the velocity data (Figure 4a) and in the geomorphological interpretation, which illustrates a long, stagnant, debris covered tongue (Figure 7). The other dataset in which a surge-front may be present is Unnamed1. This glacier is particularly interesting because the surge appears to have overridden debris (or even 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. (2014); the Kunyang

Glacier showed extensive areas of thermokarst pre-surge indicating stagnant or slow-moving ice, and the main glacier into which the Kunyang feeds (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 these surge fronts could simply be a consequence of the individual glacier geometries 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 datasets 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 dataset, although the Saxinitulu and Qiaogeli surges and previous profiles for the Khurdopin Glacier and the Gasherbrum Glacier (Figure 6) are similarly characterised. 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 build-up of fast-flowing ice is apparent (Fowler et al., 2001). Similarly, the dynamic evolution of surges observed on smaller glaciers in our dataset (Unnamed1, 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).

From the twelve velocity datasets we have now derived for Karakoram glacier surge events, there is a mix of evidence relating to the dominant trigger mechanism operating in the region (Table 3). A number of characteristics support the surges being thermally rather than hydrologically controlled: (1) the shape of the build-up, active surge and termination phases of the Karakoram surges contrast with those reported from Alaskan glaciers (e.g. Burgess et al., 2012), where hydrology is the surge control. Significantly, 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., 1987). In the Karakoram, on many glaciers the termination phase can last for years (Figure 9), suggesting in these cases the mechanisms operating are fundamentally different to those operating in Alaska. (2) The length of the build-up 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 modelled dynamics, suggesting the three-year build-up 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, observations that have been used elsewhere to 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 significantly 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 surging glaciers are hard or soft, although field observations have identified thicknesses of several metres 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 surging glaciers crossing two or more major formations (Hewitt, 1998), and possibly underlain by spatially variable geothermal heat flow (Chamberlain et al., 2012).

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 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 surge behaviours, and cannot be collectively characterised. The implication of this is that regional triggers may also not be singularly defined, and are likely to differ even on an individual glacier basis.

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 (months to years long) and do not appear to be seasonally controlled, 3. The frontal advances of some small surging glaciers can exceed 1 km over several years of surging, 4. Surge fronts are present in some Karakoram surges, but may simply reflect individual glacier geometries, 5. Uniform acceleration and deceleration across the whole glacier surface, more typically characterises these fast-flow events, 6. Maximum velocities are of the order of 2 km a^{-1} as has been reported in previous work, and 7. Surging tends to peak, and often decelerate, during summer months. Their dynamic evolution does not therefore fit neatly within either of the classically cited thermal or hydrological models of surging, suggesting factors that we still have little knowledge about (e.g. basal thermal and sedimentological conditions) are likely to be dominant controls.

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Figure 1: The Karakoram region and the location of the eight glaciers analysed in this study. Landsat background imagery © USGS, 2009+2010. Co-ordinates are given in UTM WGS84 Zone 43N. Note the image has been rotated counter-clockwise from true north.

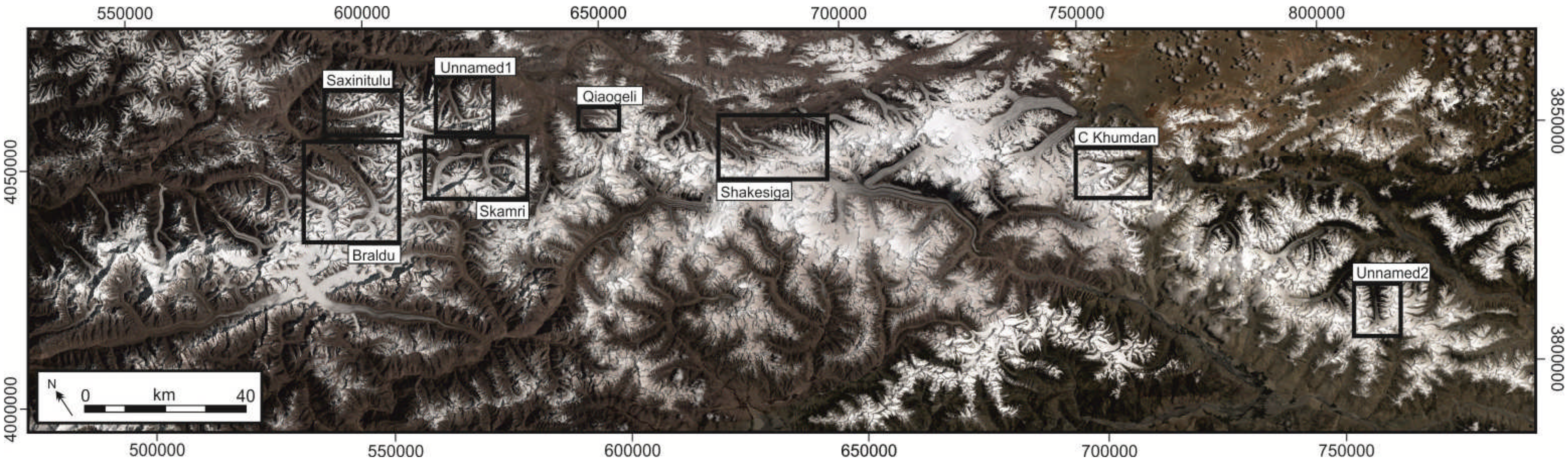

Figure 2: Detailed view of the eight glaciers and the centreline profiles used to extract velocity data (shown in Figure 4). In each case the profile is taken from the maximum terminus position reached during each glacier surge, and from the terminus moving up-glacier.

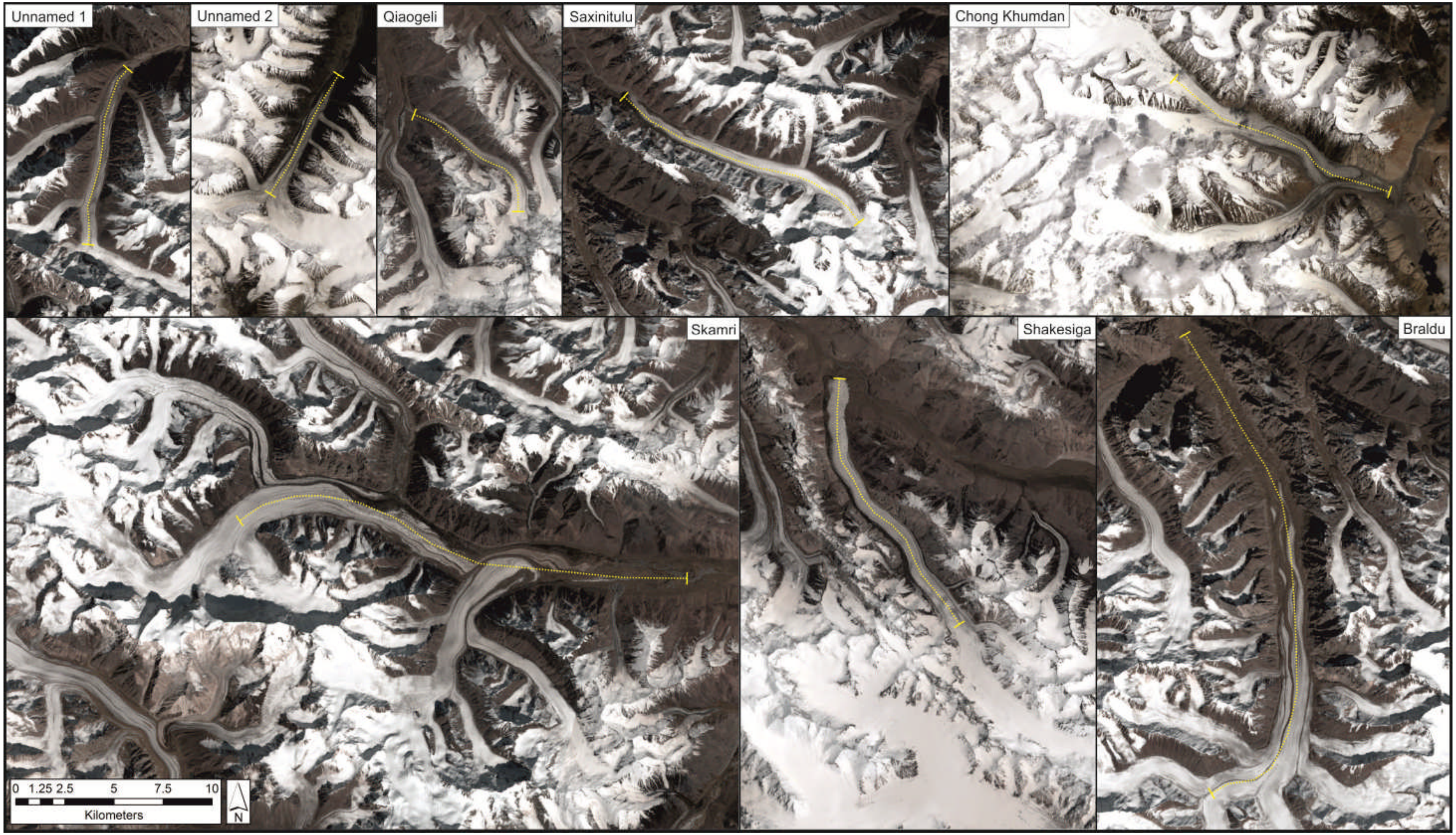

Figure 3: Selected filtered velocity fields for each of the eight glaciers.

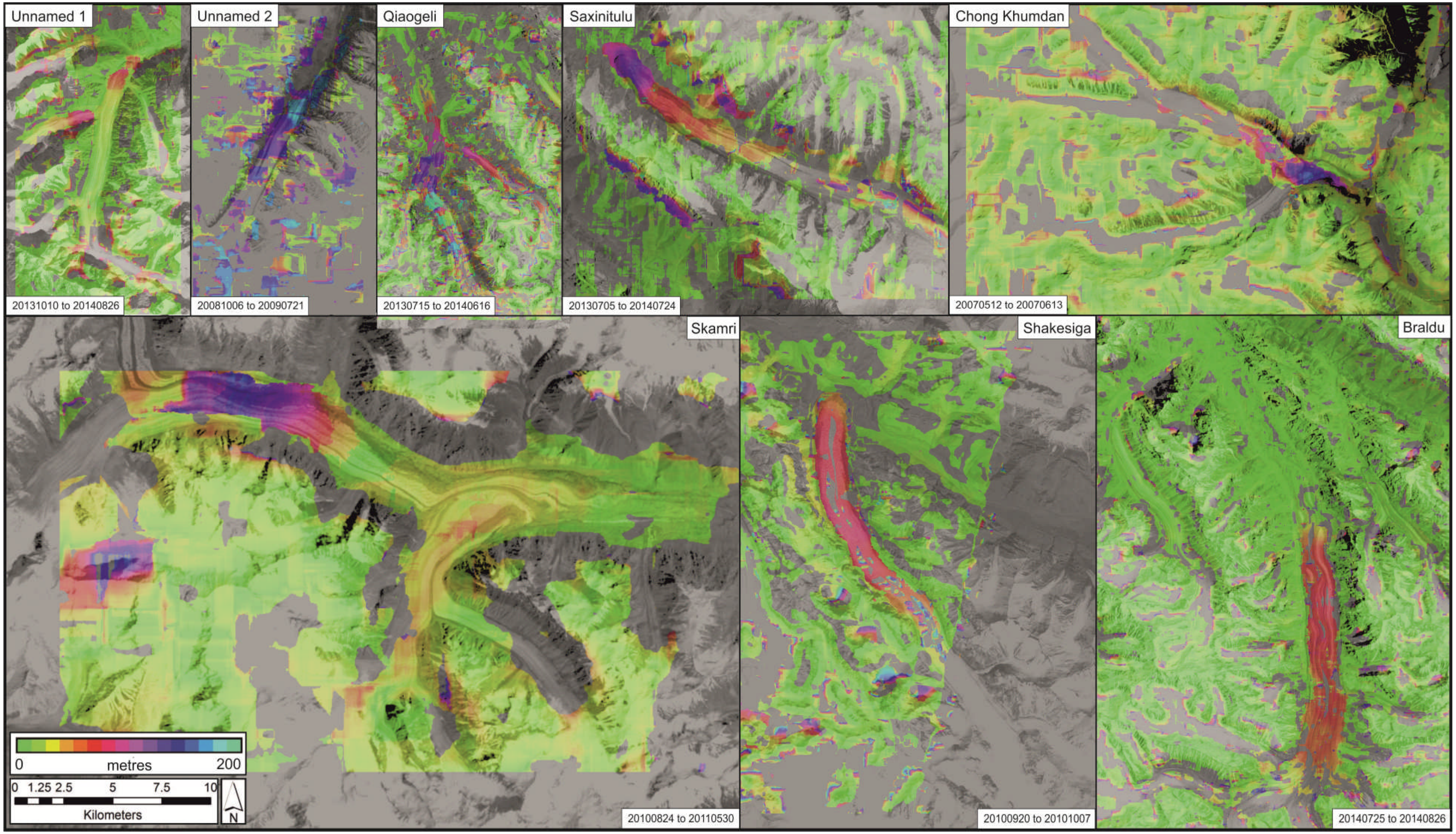

Figure 4: Centreline velocity profiles characterizing the dynamic evolution of surges on each of the eight glaciers in the study. For error estimation see Table 2. Axes scales are not directly comparable. Note that surge velocities are between one and two orders of magnitude greater than quiescent velocities in each case, and the clear down-glacier migration of a surge front in the Braldu dataset (labelled).

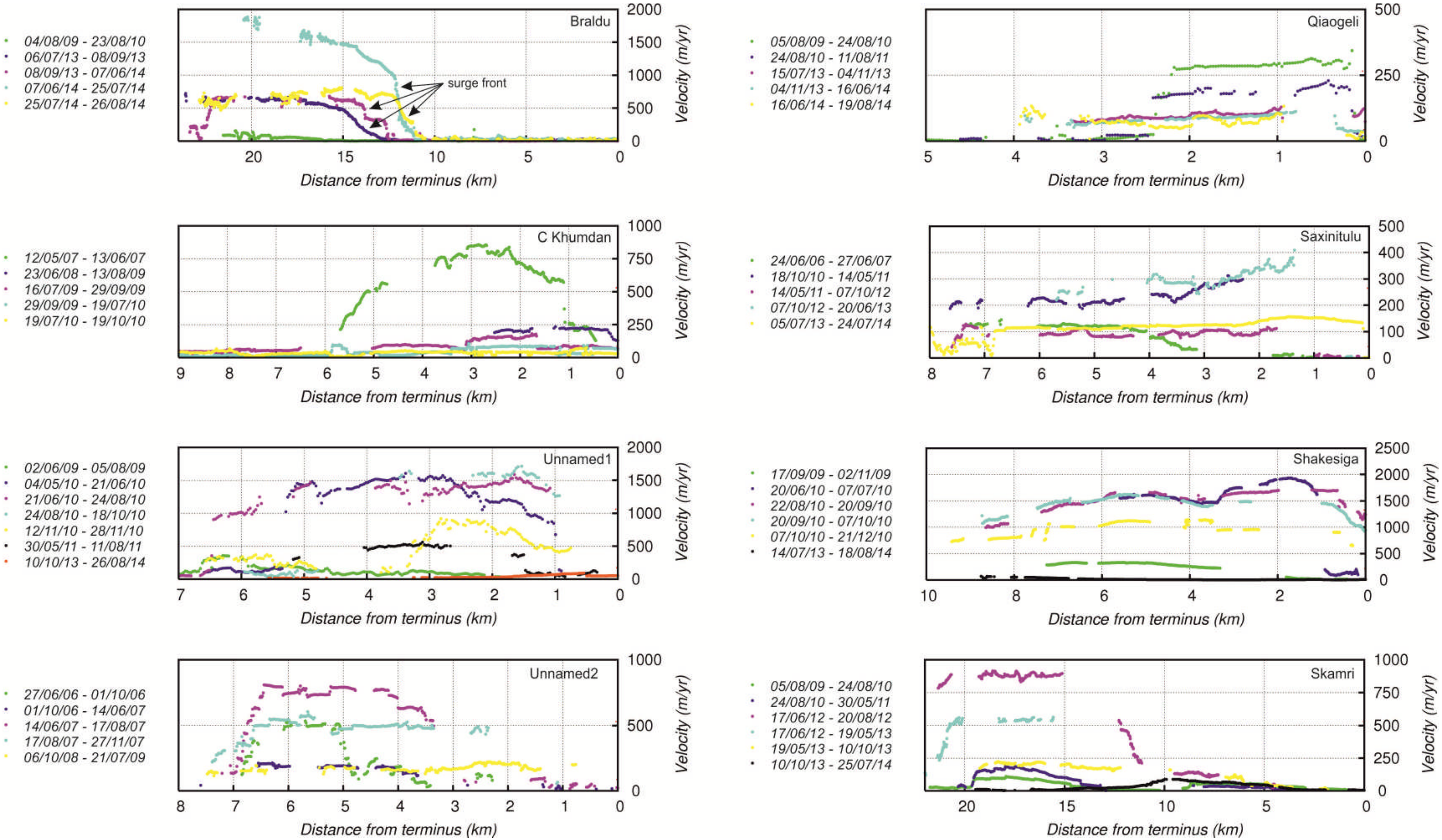

Figure 5: Before and during the surge of Saxinitulu Glacier. The surge began in 2009 and peaked in 2013. The glacier terminus is still advancing in 2015 imagery.

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).

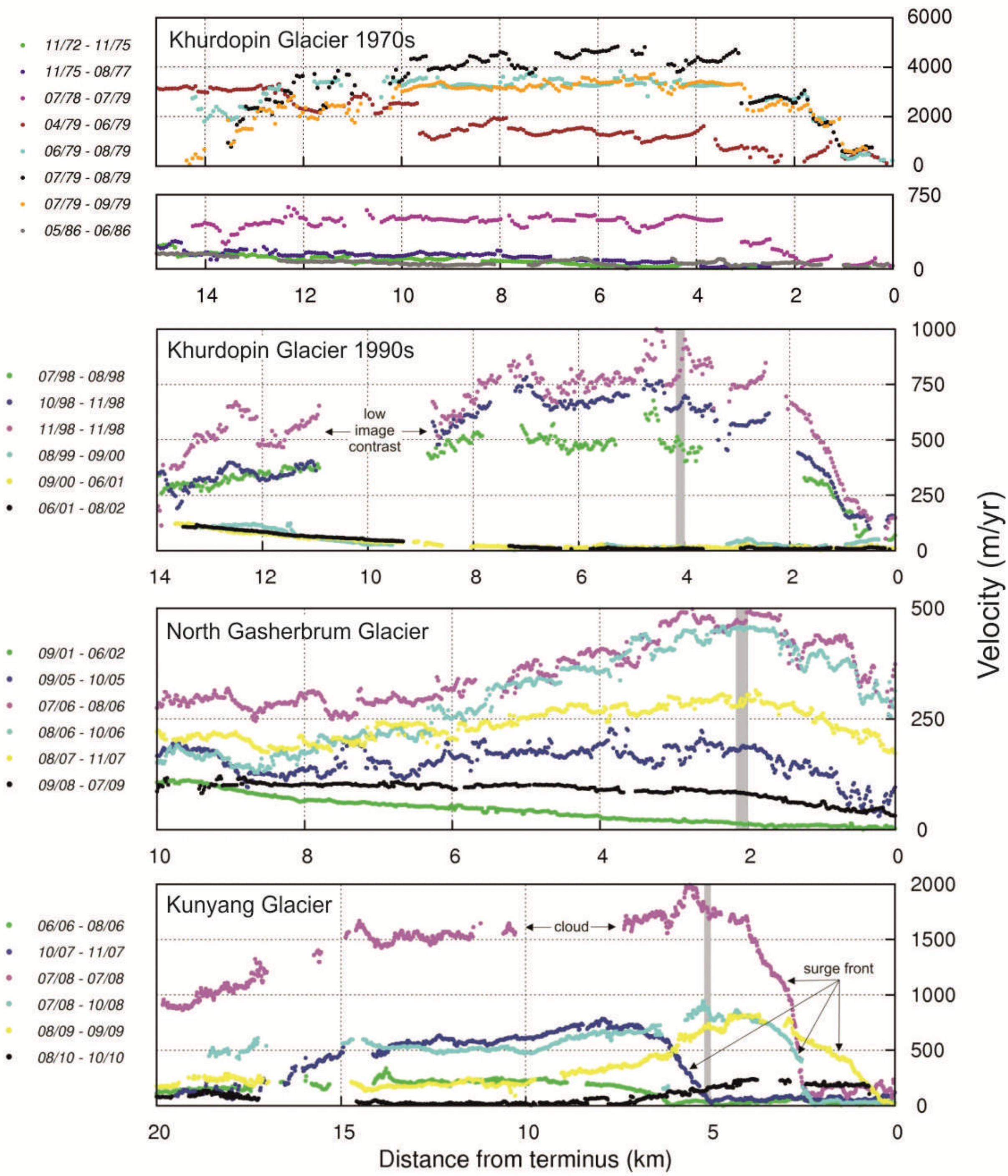

Figure 7: The geomorphic context of the Braldu surge. Black dashed lines indicate prominent surface features and their relative positions in each dataset. Grey dashed lines in the August 2014 dataset denote the relative position of each feature in the July 2013 dataset. Note the long debris covered tongue that provides a major obstacle to the down-glacier propagation of the surge front.

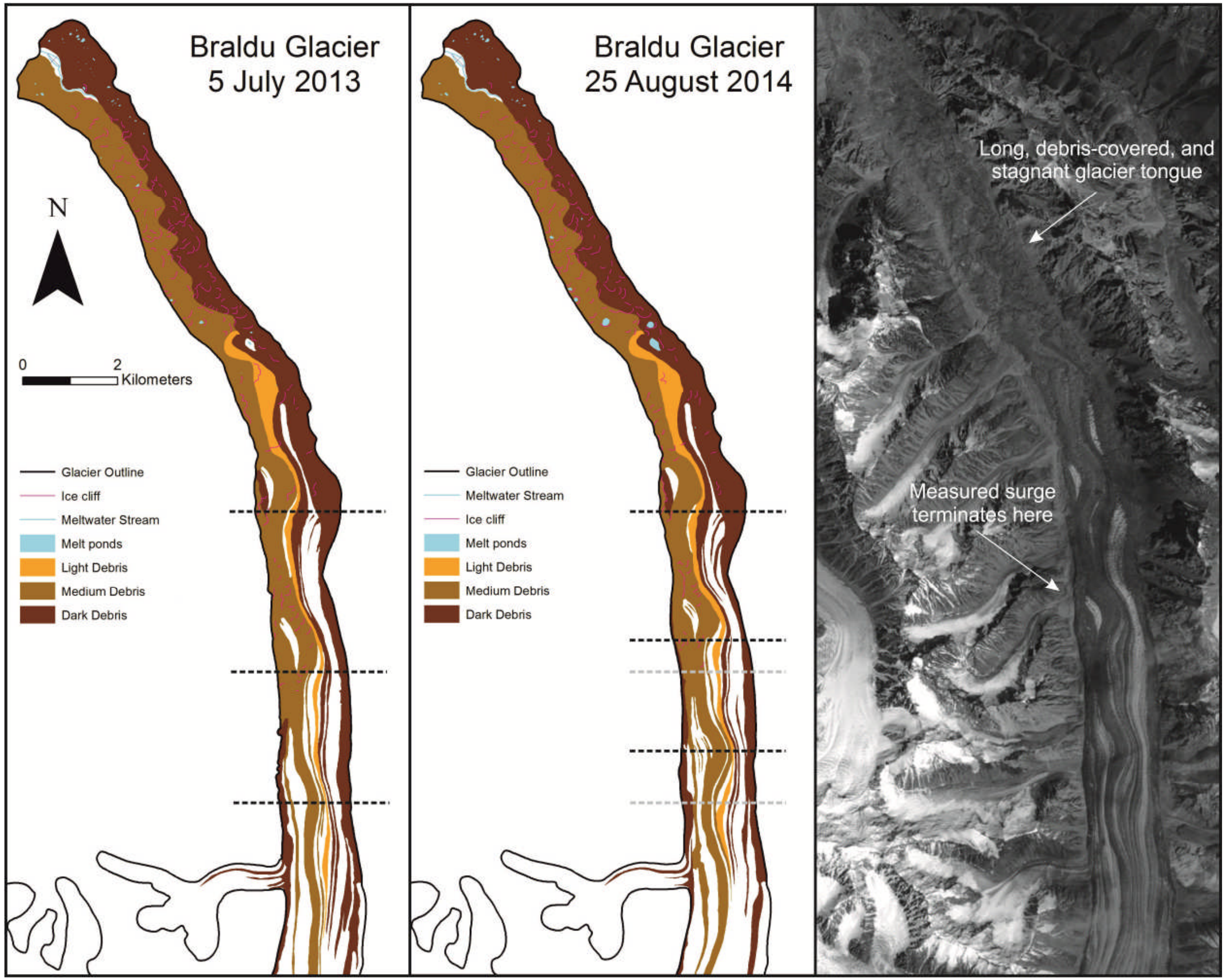

Figure 8: Evolution of the Unnamed1 surge. Note the former glacier position approximately one kilometre down-valley of the active terminus in 2009, and the way in which that ice-debris mix is overridden by the most recent surge event.

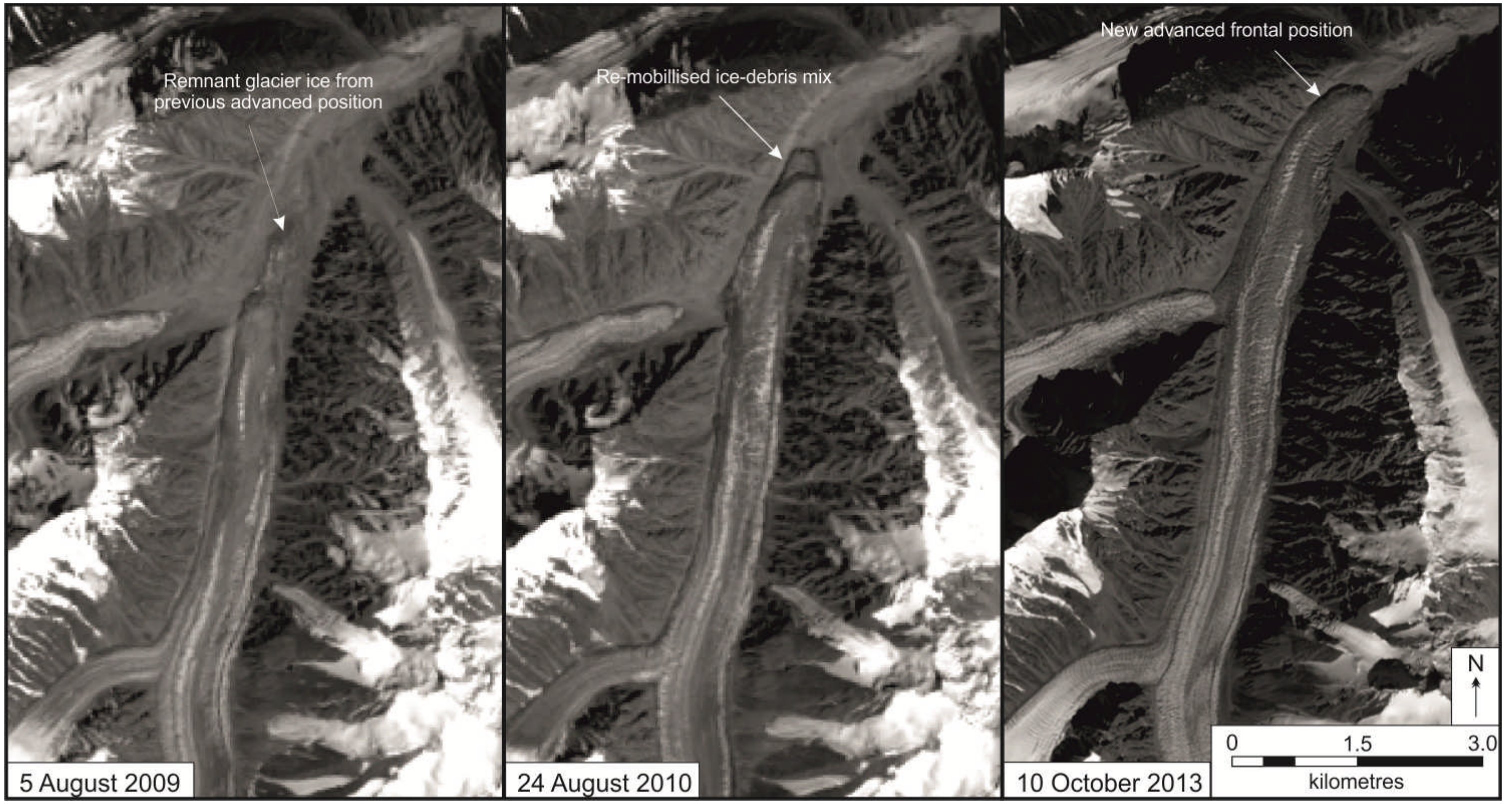

Figure 9: Surge evolution of previously measured events in the Karakoram (Quincey et al., 2011) and the Shakesiga event measured here. Note the shape of the acceleration and deceleration resembles those with a thermal control in Svalbard (Murray et al., 2003), but that the relatively short overall surge 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).

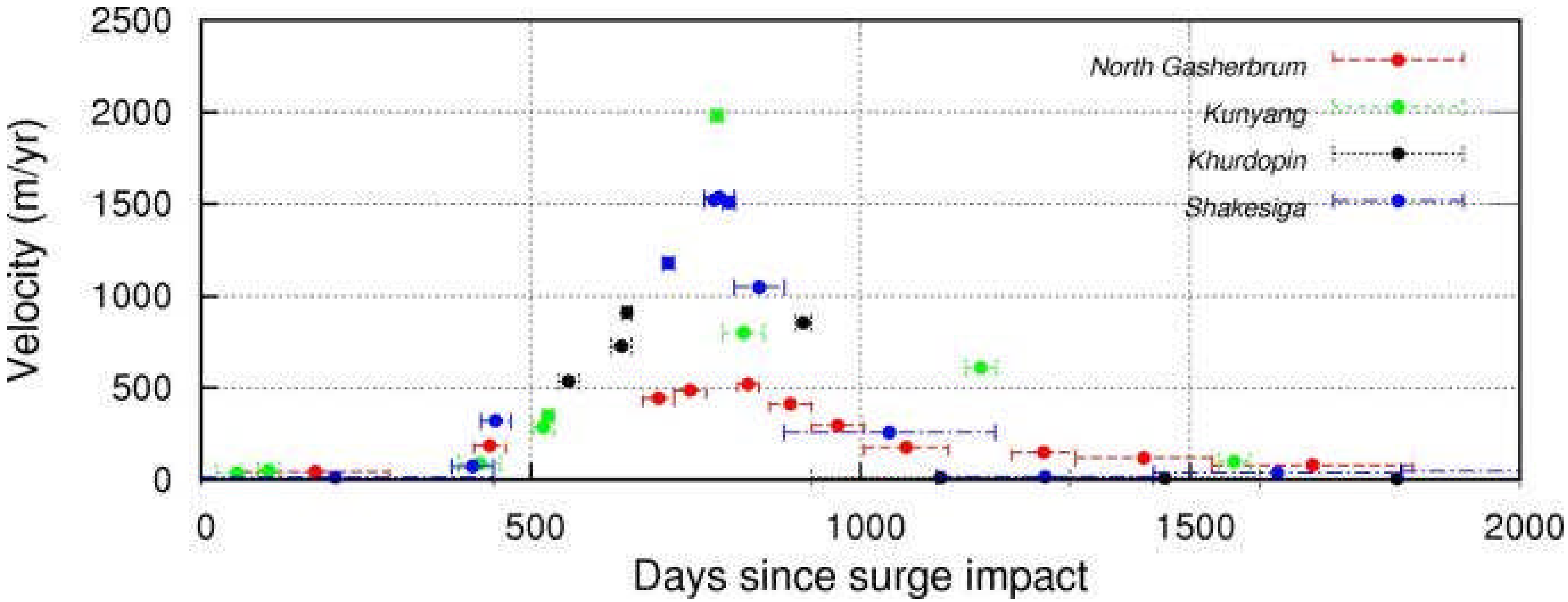

Table 1: Selected characteristics of glaciers in this study (nb. elevations and lengths are approximate values)

Glacier name	Latitude (dec deg)	Longitude (dec deg)	Max elevation (m.a.s.l.)	Min elevation (m.a.s.l.)	Length (km)	Debris covered	Aspect (degs)	Last known surge?	Reference (if applicable)
Braldu	36.143	75.865	6300	3970	34	✓	0	Unknown	Copland et al., 2011
Chong Khumdan	35.183	77.679	6370	4720	20	✓	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	×	290	Unknown	Gardelle et al., 2012
Shakesiga	35.715	76.851	7030	4420	26	×	320	Unknown	-
Unnamed1	36.178	76.202	6956	4340	14	partly	10	Unknown	-
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

Table 2: Calculated error in each of the velocity datasets presented

	Images matched	Temporal separation (days)	Sensor	Pixel resolution (m)	Calculated uncertainty (m/yr)
Braldu	20090804 to 20100823	384	TM	30	7
	20130706 to 20130908	64	ETM+	15	21
	20130908 to 20140607	272	ETM+	15	5
	20140607 to 20140725	48	OLI	15	29
	20140725 to 20140826	32	OLI	15	43
Chong Khumdan	20070512 to 20070613	32	ASTER	15	43
	20080623 to 20090813	416	TM	30	7
	20090716 to 20090929	75	AVNIR	10	12
	20090929 to 20100719	293	AVNIR	10	3
	20100719 to 20101019	92	AVNIR	10	10
Qiaogeli	20090805 to 20100824	384	TM	30	7
	20100824 to 20110811	352	TM	30	8
	20130715 to 20131104	112	ETM+	15	12
	20131104 to 20140616	224	ETM+	15	6
	20140616 to 20140819	64	OLI	15	21
Saxinitulu	20060624 to 20070627	368	ASTER	15	4
	20101018 to 20110514	208	ASTER	15	7
	20110514 to 20121007	512	ASTER	15	3
	20121007 to 20130620	256	ASTER	15	5
	20130705 to 20140724	384	ETM+	15	4
Shakesiga	20090917 to 20091102	46	AVNIR	10	20
	20100620 to 20100707	17	AVNIR	10	54
	20100822 to 20100920	29	AVNIR	10	31
	20100920 to 20101007	17	AVNIR	10	54
	20101007 to 20101221	75	AVNIR	10	12
	20130714 to 20140818	400	ETM+	15	3
Unnamed 1	20131010 to 20140826	320	ETM+	15	4
	20090602 to 20090805	64	TM	30	43
	20110530 to 20110811	73	TM	30	38
	20101112 to 20101128	16	TM	30	171
	20100824 to 20101018	55	TM	30	50
	20100621 to 20100824	64	TM	30	43
	20100504 to 20100621	48	TM	30	57
Unnamed 2	20060627 to 20061001	96	ETM+	15	14
	20061001 to 20070614	256	ETM+	15	5
	20070614 to 20070817	64	ETM+	15	21
	20070817 to 20071127	102	ETM+	15	13
	20081006 to 20090721	288	ETM+	15	5
Skamri	20090805 to 20100824	384	TM	30	7
	20100824 to 20110530	279	TM	30	10
	20120617 to 20120820	64	ASTER	15	21
	20120617 to 20130519	336	ASTER	15	4
	20130519 to 20131010	144	ASTER	15	10
	20131010 to 20140725	288	ASTER	15	5

Table 3: Surge characteristics for all twelve events in the Karakoram that have been observed with multi-temporal velocity data. 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'.

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	••••	••••