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► **To cite this version:**

Amaury Dehecq, Noel Gourmelen, Alex Gardner, Fanny Brun, Daniel Goldberg, et al.. Twenty-first century glacier slowdown driven by mass loss in High Mountain Asia. *Nature Geoscience*, 2019, 12, pp.22-27. 10.1038/s41561-018-0271-9. hal-01952548

**HAL Id: hal-01952548**

**<https://hal.science/hal-01952548>**

Submitted on 22 Apr 2024

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**Citation for published version:**

Dehecq, A, Gourmelen, N, Gardner, AS, Brun, F, Goldberg, D, Nienow, P, Berthier, E, Vincent, C, Wagnon, P & Trouve, E 2018, 'Twenty-first century glacier slowdown driven by mass loss in High Mountain Asia', *Nature Geoscience*, vol. 12, pp. 22-27. <https://doi.org/10.1038/s41561-018-0271-9>

**Digital Object Identifier (DOI):**

[10.1038/s41561-018-0271-9](https://doi.org/10.1038/s41561-018-0271-9)

**Link:**

[Link to publication record in Edinburgh Research Explorer](#)

**Document Version:**

Peer reviewed version

**Published In:**

Nature Geoscience

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# Twenty-first century glacier slowdown driven by mass loss in High Mountain Asia

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1       Glaciers in High Mountain Asia have experienced heterogeneous rates of loss  
2 since the 1970s. Yet, the associated changes in ice flow that lead to mass redistri-  
3 bution and modify the glacier sensitivity to climate are poorly constrained. Here  
4 we present observations of changes in ice flow for all glaciers in High Mountain  
5 Asia over the period 2000-2017, based on one million pairs of optical satellite im-  
6 ages. Trend analysis reveals that in nine of the eleven surveyed regions, glaciers  
7 show sustained slowdown concomitant with ice thinning. In contrast, the stable or  
8 thickening glaciers of the Karakoram and West Kunlun regions experience slightly  
9 accelerated glacier flow. Up to 94% of the variability in velocity change between  
10 regions can be explained by changes in gravitational driving stress, which in turn  
11 is largely controlled by changes in ice thickness. We conclude that, despite the  
12 complexities of individual glacier behaviour, decadal and regional changes in ice  
13 flow are largely insensitive to changes in conditions at the bed of the glacier and  
14 can be well estimated from ice thickness change and slope alone.

15       Glaciers are thinning world-wide, at an increasing rate since the turn of the 21<sup>st</sup> century [1],  
16 with a mean mass balance of  $-0.42 \text{ m w.e.yr}^{-1}$  (meter water-equivalent per year) [2]. Glaciers  
17 on the Tibetan Plateau and surrounding ranges (Figure 1), referred to as High Mountain Asia  
18 (HMA), are no exception despite regionally-contrasted evolution: some regions are experiencing  
19 close to global mean rates of mass loss, e.g. Spiti Lahaul ( $-0.37 \pm 0.09 \text{ m w.e.yr}^{-1}$ ), West Nepal  
20 ( $-0.34 \pm 0.09 \text{ m w.e.yr}^{-1}$ ) or Nyainqêntanglha ( $-0.62 \pm 0.23 \text{ m w.e.yr}^{-1}$ ) [3], whereas glaciers  
21 north-west of the Tibetan Plateau (West Kunlun Shan, Karakoram, East Pamir) are near  
22 equilibrium or slightly gaining mass [3–5]. This contrasted pattern has persisted since the  
23 1970s [6].

24       In response to these mass changes, glacier flow is expected to change, thereby affecting  
25 ice fluxes, hypsometry (ice area-altitude distribution) and glacier mass balance. However, the  
26 link between these different components and, in particular, the flow response of glaciers to mass  
27 change are poorly understood at regional scales [7]. Dynamic mass redistribution is particularly  
28 critical in regional glacier models used to estimate glacier contributions to sea-level change [8–  
29 10] and water resources [11] but is generally represented by empirical scalings, which lack a  
30 physical representation of glacier flow [12]. A few studies have attempted to model ice flow at  
31 regional scales, taking into account ice deformation [12, 13] or basal sliding [14, 15], but the  
32 justification of model choice is generally undermined by the lack of velocity observations [16].

33       Field measurements demonstrate that ice flow of land-terminating glaciers fluctuates with  
34 mass changes at decadal scales [17, 18]. Ref. [7] analysed ice velocity changes over recent  
35 decades using single satellite image pairs from 6 glacierized regions in the world. They conclude  
36 that ice flow slowed in regions with negative mass balance but found no clear relation between  
37 mass balance and velocity change. The slowdown of several land-terminating glaciers has been  
38 observed locally in HMA, concomitant with negative mass balance [19–22] but no observation  
39 of velocity changes exist at regional scales.

40       In this study we measure regional changes in the flow of HMA glaciers using systematic  
41 feature-tracking of repeat satellite images collected between 2000-2017. We discuss regional  
42 differences in velocity trends with regards to known ice thickness changes over a similar period.  
43 Finally, we estimate the contribution of changes in gravitational driving stress to the observed  
44 changes in surface velocity and discuss the best representation of ice flow in models of glacier  
45 evolution.

## 46 Interannual changes in glacier velocities

47 We derive glacier surface velocities by applying feature-tracking to 907,142 panchromatic Landsat-  
48 7 image pairs (15 m resolution) separated by less than 545 days using JPL auto-RIFT software  
49 [23] (Methods section). We generate a mean velocity field for 94% of all glaciers in HMA from  
50 an error-weighted average of all velocity fields over the period 1999-2017 at 240 m resolution.  
51 Annual velocities are obtained similarly at yearly interval for the period 2000-2017 (insufficient  
52 data was available for 1999) with glacier coverage ranging from 83 to 89% (Figure S1). Image  
53 pairs span one year on average, centred around June, with little interannual variability (Fig-  
54 ures S2 and S3). Consequently, our results are relatively insensitive to seasonal fluctuations  
55 in ice flow. The velocity uncertainty, estimated over ice-free terrain, varies with the number  
56 of available image pairs and changes in radiometric quality [24]. The median uncertainty of  
57 the annual velocity fields is around 2 m/yr, with a minimum ( $\sim 0.8$  m/yr) for the central and  
58 eastern Himalaya and a maximum (3 m/yr) for the Tibetan Plateau (Figure S4). Examples of  
59 velocity maps are shown on Figure S5 (a,c,e).

60 To extract regional velocity trends, we conduct our analysis on glacier areas with surface  
61 velocities that significantly exceed the estimated uncertainty. We select pixels with a mean  
62 velocity greater than 5 m/yr over glaciers larger than 5 km<sup>2</sup> (areas from ref. [25]). Accumulation  
63 zones have larger measurement uncertainties due to low image contrast and have experienced  
64 little elevation change [3]. For this reason, we restrict our analysis to the lower half of each  
65 glacier, which approximately represents the ablation zone. Glaciers known to experience surges  
66 [26] (surge-type) are included in the general analysis but their regional response is also quantified  
67 and discussed separately. Our observations are uniformly distributed across altitude in the  
68 ablation zone and glacier size (above 5 km<sup>2</sup>) and are therefore representative of the diversity  
69 of glaciers in HMA (Supplementary section 1).

70 To characterize regional changes in ice flow, we examine anomalies in annual velocity for  
71 each region. We define the velocity anomaly as the vector difference between the annual velocity  
72 and the mean velocity, projected to the orientation of the mean velocity (Methods). This scalar  
73 variable is positive if the ice flow accelerated along a flow line and negative if it slowed down.  
74 This approach ensures that the uncertainty in velocity change is symmetrically centred on zero  
75 (Figure S8) as opposed to simply differencing the velocity magnitude (Figure S9). We calculate,  
76 for each year and 11 subregions in HMA, the median anomaly over pixels with observations in

77 all years, and compute a linear trend over the period 2000-2017 (Methods).

78 The results (Figure 1) show that the largest velocity changes (slowdown) occur for glaciers  
79 in Nyainqêntanglha ( $-37.2 \pm 1.1$  %/decade) and Spiti Lahaul ( $-34.3 \pm 4.5$  %/dec). Smaller  
80 but significant slowdowns are observed along the Himalayan range with decreasing amplitude  
81 towards the East: West Nepal ( $-21.0 \pm 2.3$  %/dec), East Nepal ( $-17.0 \pm 1.0$  %/dec) and Bhutan  
82 ( $-14.5 \pm 1.3$  %/dec). Contrasted trends are observed in the north-western regions, with negative  
83 trends in the Hindu-Kush ( $-9.8 \pm 2.9$  %/dec), Pamir ( $-9.4 \pm 1.6$  %/dec) and Tien Shan ( $-6.4$   
84  $\pm 1.0$  %/dec) while a small but significant speed-up is observed for the Karakoram ( $3.6 \pm 1.2$   
85 %/dec) and West Kunlun ( $4.0 \pm 2.1$  %/dec). Finally, the inner Tibetan Plateau (TP) displays  
86 a negative trend ( $-8.2 \pm 2.3$  %/dec). Very few observations of glacier velocity changes exist  
87 in HMA for validation, but our results show good agreement with both field [20] and remote  
88 sensing [22] observations (Supplementary section 2).

89 Our results reveal that changes in velocity are not always monotonic over the study period.  
90 Most regions in the north-west (Pamir, Hindu-Kush, Spiti Lahaul and West Nepal) experienced  
91 a pronounced slowdown until 2005-2008 with more stable conditions since. On the contrary,  
92 East Nepal and Nyainqêntanglha experienced a steady and continuous slowdown while Bhutan  
93 experienced a slight increase in its rate of slowdown after 2008. These patterns are consistent  
94 with glacier mass balance and elevation change trends [27]. It is noteworthy that the strongest  
95 trends are generally observed over 2003-2008, coinciding with the period of observations of the  
96 satellite altimeter ICESat, suggesting that elevation changes derived from ICESat are poten-  
97 tially more negative than the longer term trend [3].

98 Our analysis focuses on results determined from a single sensor (Landsat 7) due to biases  
99 that we identified between velocities derived from different Landsat missions (Supplementary  
100 section 3). However, trends estimated between 1988 and 2017 with over 2 millions image  
101 pairs from the Landsat missions 5-8, and accounting for inter-mission biases, lead to similar  
102 results despite larger uncertainties, except for a break in trend observed for Spiti Lahaul and  
103 Nyainqêntanglha around year 2000 (Figure S12). This is consistent with stable conditions  
104 observed in Spiti Lahaul for the 1990s [28, 29].

## 105 Correlation between regional velocity trends and mass bal- 106 ance

107 Trends in velocity anomalies are calculated for each 240-m pixel over the period 2000-2016  
108 (Methods section) to match the observation period of glacier thickness change [3]. Examples  
109 of velocity trend maps are shown on Figure S5. The results are presented as a median velocity  
110 trend on a  $1^\circ \times 1^\circ$  grid (Figure 2a) next to rates of elevation change (Figure 2b). The similarity  
111 between patterns of velocity and thickness change, the latter being largely driven by differences  
112 in mass balance sensitivity to temperature [30] and different climatic conditions [31], suggests  
113 that the spatial variability in velocity change is also influenced by regional differences in climate  
114 and glacier sensitivity to temperature. Slowdown along the Himalayan range, Nyainqêntanglha  
115 or Tien Shan, for example, is associated with ice thinning, whilst stable or increased glacier flow  
116 is observed along with stable to positive mass balance around the Tibetan Plateau and Tarim  
117 basin (the so-called "Karakoram anomaly"). Some trends differ however, notably the speedup  
118 observed in western Tien Shan in a region of negative mass balance and areas of slowdown in  
119 West Kunlun, a region of positive mass balance. The regional velocity trend derived for these  
120 regions is particularly sensitive to the selected area and the discrepancies are likely related  
121 to the incomplete spatial sampling of the velocities, or to the large flow variability of these  
122 regions, caused by surge activity for instance. At regional scales, changes in glacier velocity  
123 and glacier-wide mass balance are strongly correlated (Figure 3a,  $R^2=0.76$ ). This relationship  
124 implies that surface velocity change can be used as a proxy for regional glacier evolution in  
125 areas and during periods where regional glacier mass balance is not available. This possibility  
126 is especially interesting since surface velocity is more easily obtained from remote-sensing than  
127 elevation measurements.

## 128 Ice dynamical response to thickness change

129 Ice flow is primarily controlled by the driving stress (horizontal component of the ice weight  
130 per unit area), which causes ice deformation (creep) and sliding over or deformation of the  
131 bed [32]. Observations (ref. [32] section 8.3) have shown that surface velocity due to creep  
132 is a function of the glacier thickness and driving stress. Basal velocity on the other hand is  
133 poorly constrained due to complexities at the glacier bed (bed roughness, type of bed) and is



134 often represented by a power-law of the driving stress [33, 34]. In these theoretical frameworks,  
 135 the surface velocity  $U_s$  observed by remote sensing, the sum of both contributions, is therefore  
 136 expected to respond instantaneously to a change in driving stress. Field-based studies on the  
 137 other hand have observed a relationship between mass balance and ice flow changes with a lag  
 138 of 1-3 years, suggested to be the time taken to diffusively propagate a change in mass load  
 139 [17, 18].

140 Here, we assume that surface velocity can be represented by the relationship (see Methods  
 141 and Supplementary section 5):

$$U_s = C\tau^m \quad (1)$$

142 where  $C$  and  $m$  are unknown parameters and  $\tau$  is the driving stress. By allowing for differing  
 143 values of the exponent  $m$ , this relationship encompasses flow due to both ice deformation and  
 144 basal sliding (see Methods).  $C$  is likely to vary spatially and depends on local parameters such  
 145 as ice rheology, valley shape and bed roughness, while  $m$  is related to the processes leading to  
 146 ice flow. We further assume that  $C$  and  $m$  do not vary significantly with time over the study  
 147 period. In these conditions, a change in velocity  $\delta U_s$  is related to a change in driving stress  $\delta\tau$   
 148 by:

$$\log\left(1 + \frac{\delta U_s}{U_s}\right) = m \log\left(1 + \frac{\delta\tau}{\tau}\right) \quad (2)$$

149 with  $m$  to be determined.

150 To test these hypotheses, we calculate the change in driving stress, associated with the  
 151 changes in thickness observed by ref [3], along glacier centre flow lines for the period 2000-  
 152 2016. Measurements of ice thickness are required to calculate the exact driving stress (equation  
 153 10), but are unavailable across the whole HMA. We therefore use modelled thickness estimates  
 154 that have uncertainties of  $\sim 25\%$  [35, 36]. We use the ice surface elevation from the Shuttle  
 155 Radar Topography Mission (SRTM) version 3 [37] for year 2000 and from an application of  
 156 elevation change rates over the period 2000-2016 [3] for year 2016. Ice thickness and elevation  
 157 are extracted along glacier center flow lines at 50-m spacing to calculate a relative change in  
 158 thickness and driving stress between 2000 and 2016. We perform the calculations for 2894  
 159 glaciers larger than 5 km<sup>2</sup> and calculate a median driving stress change in ablation zones for  
 160 each subregion, that we compare with median velocity changes calculated over the same points.

161 To identify a possible lag of a few years between driving stress and velocity change, we compare  
162 the driving stress change over 2000-2016 with the observed velocity anomaly trends for three  
163 periods: 2000-2016 (instantaneous response), 2001-2017 (1 year lag) and 2003-2017 ( $\sim 3$  year  
164 lag). Larger lags are not considered due to the lack of observations after 2017, which increases  
165 uncertainties in later trends.

166 Our results show that changes in driving stress can explain up to 94% of the inter-regional  
167 variability in the observed velocity change with a 3-year lag (Figure 3b,  $R^2=0.94$ ). We also  
168 observe that the strength of the correlation is improved with a 3-year lag as opposed to a  
169 1-year lag ( $R^2=0.85$ ) or no lag ( $R^2=0.75$ ) (Figure S17). Possible explanations for this lag are  
170 the diffusive propagation of the thickness change, or adjustment of the bed and subglacial  
171 environment to thickness change, that cause a delay in the velocity response [17]. A least-  
172 squares regression indicates that the velocity change is best represented by the power  $m=4.0$   
173 (68% confidence interval [3.4-4.7]) of the change in driving stress.

174 The change in driving stress is a combination of change in thickness and slope (Eq. 10).  
175 Glacier thinning is generally more pronounced at lower elevations [3], causing an increase in  
176 slope, that in turn counteracts the reduction in driving stress caused by the thinning. Our  
177 results show that the change in driving stress obtained by taking into account the change in  
178 thickness and slope is reduced by 15% as compared to accounting for thickness alone (Figure  
179 S18). The change in slope indeed offsets the impact of the thinning at lower elevations, but  
180 thickness change remains the main contributor to the change in driving stress.

181 Many glaciers in HMA, mostly located in the Karakoram, Pamir, West Kunlun and Tien  
182 Shan [26], experience surges, i.e velocity fluctuations primarily driven by internal glacier insta-  
183 bilities as opposed to climate (Ref. [32] chap. 12). It is important to determine whether such  
184 glaciers must be considered separately for future projections. Surge-type glaciers are identified  
185 using previous studies [5, 38, 26, 39] and from the data generated as part of this study (Supple-  
186 mentary section 4). Our results do not differ significantly when surge-type glaciers are excluded,  
187 in particular the annual velocity anomaly time-series (Figure S14) or the velocity trend map  
188 (Figure S15). The power-law relationship between driving stress and velocity change remains  
189 unaltered with surge-type glaciers both included (Figure 3b, black-edge dots) or excluded (Fig-  
190 ure 3b, grey-edge dots,  $R^2=0.95$ ). The fact that surge-type glaciers have a similar regional  
191 average response as other glaciers is likely due to the heterogeneity in surge characteristics  
192 (onset, duration) within a region [39], which tends to average out over sufficiently large spatial

193 and temporal scales. As a consequence, surging behaviour does not need to be considered to  
194 correctly estimate the average flow response of glaciers at regional scales.

## 195 Implications for regional glacier models

196 Our findings have important consequences for understanding, and thus modelling, glacier re-  
197 sponse to environmental forcing. Our results show that the main driver of decadal and regional  
198 velocity change is the change in driving stress (Figure 3b), primarily attributable to changes in  
199 thickness (Figure S18). This is supported by ref. [40] who showed that changes in basal and  
200 surface velocity of the Argentière glacier, French Alps, are driven by thickness change. This  
201 implies that ice flow response to external forcing over decadal time scales can be estimated  
202 from the glacier’s slope and thickness alone, which are pre-requisites for any glacier flow model.  
203 More complex factors such as basal conditions, ice rheology or lateral drag associated with  
204 thinning or changing melt regimes, and largely unknown at regional scales, play only a minor  
205 role on decadal flow variability and for the range of change in driving stress observed here. It  
206 is important to note however that driving stress alone does not explain the large inter-glacier  
207 variability in our observations (Figure S19). A possible explanation is the uncertainty in indi-  
208 vidual glacier thicknesses used to calculate the driving stress [36]. Another possible explanation  
209 is that changes in subglacial water pressure associated with inputs of surface meltwater, known  
210 to play a significant role in driving seasonal fluctuations in surface velocity [41, 40] have a larger  
211 contribution at the glacier scale, as opposed to the regional scale.

212 Another uncertainty in glacier modelling is the fraction of basal sliding, known to be impor-  
213 tant in temperate and polythermal valley glaciers [42, 43, 40]. A change in driving stress will  
214 impact both basal sliding and ice deformation. Our results suggest that surface velocity evolves  
215 with the power  $m = 4$  of the driving stress. This is consistent with sliding theories incorporat-  
216 ing cavitation (ice-bed decoupling in the lee of obstacles when the subglacial pressure is high)  
217 [44] leading to an exponent  $m$  larger than 3 (see Methods). Furthermore, because changes  
218 in thickness and driving stress are very strongly correlated (Figure S18), creep velocity is a  
219 function of the fourth power of the driving stress, also compatible with our results (see Meth-  
220 ods). This implies that both contributions evolve similarly with the driving stress and their  
221 relative contribution remain the same even as driving stress varies. It follows that we cannot  
222 separate the contribution of changes in basal sliding and creep velocity to the surface velocity

223 change. More importantly, it also means that surface glacier velocity change can be modelled  
224 and parametrised without *a-priori* knowledge or assumptions regarding the fractional contri-  
225 bution of basal sliding. It must be noted however that the value retrieved for  $m$  is strongly  
226 conditioned by the uncertainty in current thickness estimates. An error in the exponent  $m$   
227 would lead to an error in ice transport to lower elevations and ice melt (an underestimation of  
228  $m$  would lead to an overestimate of mass transport and melt in a thinning scenario, see Supple-  
229 mentary 6), with large implications for future estimates of glacier mass changes. However, the  
230 complex relationship between ice flow, ice redistribution and mass balance makes it difficult  
231 to estimate the impact on the final mass budget. We therefore encourage studies combining  
232 observations of decadal glacier flow changes and glacier models to better constrain and reduce  
233 uncertainties in glacier dynamics.

234

235 In this study, we documented the evolution of surface velocities in the ablation zone of  
236 glaciers larger than 5 km<sup>2</sup> in High Mountain Asia between 2000-2017, providing an unprece-  
237 dented and detailed picture of glacier flow response to recent climate change. Our results  
238 reveal regionally-heterogeneous trends in surface velocity that parallel changes in ice thickness.  
239 Regions of rapid thinning show the largest rates of slowdown (Nyainqêntanglha, Spiti Lahaul)  
240 while regions near balance or gaining mass have experienced a slight speedup (Karakoram, West  
241 Kunlun). The strong relationship between regional glacier mass balance and velocity changes  
242 reveals a quasi-instantaneous response of ice flow to climate forcing and suggests that surface  
243 velocity can be used as a proxy for glacier state at decadal scales. Analysis along glacier flow  
244 lines shows that, at regional scales, 94% of the observed velocity changes can be explained by  
245 changes in driving stress, the latter being primarily controlled by changes in ice thickness. Our  
246 results suggest that changes in glacier flow in response to mass changes can be estimated in  
247 regional glacier models from ice thickness and slope alone, despite poorly constrained basal  
248 conditions and rates of basal sliding. These conclusions emphasize the important role played  
249 by ice dynamics in the glaciers response to environmental forcing and will lead to improved  
250 modelling of climate-glacier feedbacks and estimates of glacier contributions to hydrology and  
251 sea-level change.

## 252 Methods

### 253 Surface velocity

254 The JPL autonomous Repeat Image Feature Tracking (auto-RIFT version 0.9) processing  
255 scheme [23] was applied to all Landsat 4, 5, 7 and 8 Collection 1 LT1 images acquired over  
256 HMA between 1985 and 2017 with 60% cloud cover or less, as indicated in the image metadata.  
257 The images are pre-processed using a 5 by 5 Wallis operator to normalize for local variability  
258 in image radiance caused by shadows, topography and sun angle. For Landsat 4 and 5, along-  
259 track artefacts [45] are removed using Fourier filtering and a Principal Component Analysis  
260 of bands 1 to 4 is used, whereas for Landsat 7 and 8 panchromatic (Band 8) images are used  
261 (15 m pixel size). Missing data in Landsat 7 images introduced after the Scan Line Corrector  
262 failure (SLC-off) are filled with random data so that they do not contribute to the amplitude of  
263 the correlation peak. Pre-processed image pairs were searched for matching features by finding  
264 local normalized cross correlation (NCC) maxima at sub-pixel resolution by oversampling the  
265 correlation surface by a factor of 16 using a Gaussian kernel and identifying the location of  
266 maximum correlation. The use of a Gaussian kernel greatly reduces the sensitivity of subpixel  
267 displacement estimates to "pixel-locking" [46]. A sparse (1/4 of full search) NCC search is first  
268 used to determine areas of coherent correlation between image pairs. Results from the sparse  
269 search guide a dense search with search centres spaced such that there is no overlap between  
270 adjacent template windows. For HMA, image pixels located within a 2 km buffer of glacier  
271 surfaces were searched with a 240 m by 240 m search window. Image pixels located more  
272 than 2 km from a glacier were searched with a 480 m by 480 m search window with areas of  
273 unsuccessful retrievals searched with a 960 m by 960 m window.

274 Image geometry between image pairs is highly stable, but images suffer from x and y geolo-  
275 cation errors of typically  $\sim 15$  m. To correct for geolocation errors the component velocities are  
276 tied to stable surface wherein the median of each velocity component ( $V_x$ ,  $V_y$ ) is set to zero over  
277 non-glacier surface. Velocity fields were also contaminated by match blunders (e.g. matching  
278 along shadow edges or of surfaces obscured by cloud in one of the two images). Component  
279 velocities that deviate by more than 3 times the interquartile range from the median of all  
280 co-located pixels are assumed to be gross outliers and are removed. The uncertainty of each  
281 image-pair velocity field is set equal to the standard deviation in component velocities measured

282 over stable surface.

283 Annual velocity maps are created by taking the error-weighted average of all image-pair  
284 velocity fields having a centre-date that fall within that calendar year. A mean velocity field  
285 ( $\vec{V}_0$ ) is then created by taking the error weighted average of all annual velocity maps. The  
286 uncertainty of the merged velocities is estimated on a pixel basis by propagating the uncertainty  
287 of each measurement:

$$\sigma_X = \sqrt{\frac{\sum \sigma_{i,X}^2}{N}} \quad (3)$$

288 Where X denotes the component x or y,  $\sigma_i$  is the uncertainty of each individual velocity field as  
289 estimated from the stable areas and N is the number of observations contributing to the weighted  
290 average. An effective date and pair time span are estimated for each pixel as a weighted average  
291 of the individual pairs' date and time span. Using this approach, we calculated yearly velocity  
292 maps from 1985 to 2017 that were derived from 2,287,223 unique image pairs (Landsat 4: 367,  
293 Landsat 5: 836,616, Landsat 7: 907,142, Landsat 8: 543,098). For our analysis, we excluded  
294 velocity estimates with large uncertainties, i.e. where  $\sigma = \sqrt{\sigma_x^2 + \sigma_y^2} > 5$  m/yr and  $N < 5$ .

## 295 Velocity change

296 We estimate the velocity change compared to the mean velocity  $\vec{V}_0$ . We define the velocity  
297 anomaly as the value of the difference vector  $\vec{V}_t - \vec{V}_0$  projected on the mean velocity vector:

$$dv = \frac{(\vec{V}_t - \vec{V}_0) \cdot \vec{V}_0}{\|\vec{V}_0\|} = \frac{(V_{x,t} - V_{x,0}) \cdot V_{x,0} + (V_{y,t} - V_{y,0}) \cdot V_{y,0}}{\|\vec{V}_0\|} \quad (4)$$

298 The difference in velocity magnitude is typically used to characterize velocity change [7, 47, 48,  
299 22]. However, if each component of the velocity can be considered as following a symmetrical  
300 distribution, the distribution of the velocity magnitude, by definition, is skewed towards positive  
301 values with a non-zero mean. In the case of normally distributed noise, the velocity magnitude  
302 follows a Rice distribution that has a biased mean [49]. This bias decreases with the velocity  
303 magnitude and increases with the standard deviation of the velocity components (noise). A  
304 consequence of this bias is an apparent negative velocity trend in slow-moving areas when  
305 estimating changes between the earlier Landsat missions with a higher noise and the newer  
306 mission with a reduced noise (Figure S9). This bias affects velocity trends in areas where  
307 the velocity is not significantly larger than the noise. The proposed velocity anomaly has the

308 advantage of having a noise that is symmetrically distributed around 0 that will not introduce  
309 a bias in the mean value (Figure S8), even for slow-moving areas.

310 *Region of Interest.* We restrict our analysis to the relatively fast moving part (mean velocity  
311 greater than 5 m/yr) of the ablation area of glaciers larger than 5 km<sup>2</sup>. Glaciers smaller  
312 than several square kilometres tend to have velocities below our uncertainty threshold, few  
313 measurements, and narrow tongues of width similar to the correlation window, which highly  
314 decreases the confidence in these measurements. The ablation zone is approximated as all points  
315 located below the glacier median elevation  $z < (z_{max} + z_{min})/2$  where  $z$  is the pixel elevation  
316 and  $z_{min}$  and  $z_{max}$  are the minimum and maximum altitude of the glacier to which the pixel  
317 belongs.  $z_{min}$  and  $z_{max}$  are extracted from the RGI 6.0 inventory [25] and  $z$  is extracted from  
318 the Shuttle Radar Topography Mission (SRTM) topography version 3 [37]. These points are  
319 later referred as the Region of Interest (RoI).

320 *Glaciers surges.* We exclude glaciers with reported surge activity for parts of the analysis.  
321 We use inventories from previous studies [5, 38, 26, 39] (Supplementary section 4). We also  
322 exclude glaciers that were not identified as surging in those inventories but display a behaviour  
323 typical of surge events (temporally and spatially limited speed-up, slowdown in an upper zone  
324 and acceleration at the tongue or reverse, thinning in an upper zone and thickening of the  
325 tongue or reverse). The outlines of surging glaciers are provided as supplementary data.

326 *Velocity anomaly time series.* To calculate the temporal evolution of the velocity anomaly  
327 for a given region, it is necessary to calculate statistics on pixels with observations for all years.  
328 However, as some years have lower spatial coverage, there is a compromise to be made between  
329 temporal and spatial coverage. The mask of common pixels is estimated as follows. The  
330 intersection of all valid pixels for all selected years is computed. If the coverage of the common  
331 mask is less than 25% of the RoI, the year with least coverage is excluded and the previous steps  
332 are repeated. As a result, years that do not meet the coverage criteria are excluded. Finally,  
333 for each region, the median and interquartile range of the velocity anomalies on the common  
334 mask are calculated. A trend in the regional velocity anomalies is calculated with uncertainty  
335 for the period 2000-2017 following the same methodology as below.

## 336 Velocity trends

337 We calculate a trend in velocity anomalies over the study period 2000-2016 for each 240 m by  
338 240 m pixel of the annual velocity maps using a linear regression:

$$dv(x, y) = a(x, y) \times t + b(x, y) \quad (5)$$

339 Where  $dv$  is the velocity anomaly for the pixel located at position  $(x, y)$ ,  $t$  is the year of observa-  
340 tion,  $a$  and  $b$  are the parameters of the linear regression estimated at each pixel. To account for  
341 outliers, we perform the regression iteratively by removing observations with residuals larger  
342 than 3 standard deviations. The standard error  $\sigma_a$  (resp.  $\sigma_b$ ) of the parameters  $a$  (resp.  $b$ )  
343 are estimated from the regression covariance matrix. We interpolate the velocity for year 2000  
344 from the linear regression parameters:

$$V_{2000} = V_0 + a \times 2000 + b \quad (6)$$

345 and compute a velocity change relative to year 2000 (percentage change per decade) as:

$$ddv = a/V_{2000} \times 10 \quad (7)$$

346 We estimate the uncertainty in the velocity trend using a Monte-Carlo method by randomly  
347 drawing the first and last velocities of the study period from a Gaussian distribution determined  
348 from the regression uncertainty and then calculating the associated velocity change  $ddv$ . We  
349 repeat the operation 200 times and calculate the standard deviation of the distribution  $\sigma_{ddv}$ .  
350 Finally, we exclude all pixels with observations in less than 50% of years or  $\sigma_{ddv} > 30\%/dec$ .

351 We generate the regional map of velocity trends by extracting the median, standard devia-  
352 tion  $\sigma_{1^\circ}$  and number of observations  $N_{1^\circ}$  of the velocity trend for  $1^\circ \times 1^\circ$  bounding boxes. The  
353 standard error is calculated as:

$$\epsilon_{1^\circ} = \frac{\sigma_{1^\circ}}{\sqrt{N_{1^\circ}}} \quad (8)$$

## 354 Impact of the driving stress

355 Ice surface velocity is taken as:

$$U_s = C\tau^m \quad (9)$$

356 where  $C$  and  $m$  are unknown parameters and  $\tau$  is the driving stress (assumed equal to the basal  
357 stress), defined as:

$$\tau(x) = \rho g H(x) \frac{\partial S}{\partial x}(x) \quad (10)$$



358 with  $\rho$  the ice density,  $g$  the gravitational acceleration,  $H(x)$  the ice thickness and  $S(x)$  the ice  
 359 surface at the position  $x$  along a given flow line. In general,  $C$  and  $m$  are poorly-constrained and  
 360 likely to vary spatially. The only hypothesis we make in our analysis regarding these parameters  
 361 are that they do not change over the time interval of interest (2000-2016). A change in driving  
 362 stress  $\delta\tau$  hence induces a change in velocity according to:

$$U_s + \delta U_s = C(\tau + \delta\tau)^m \quad (11)$$

363 Which can be rewritten as:

$$1 + \frac{\delta U_s}{U_s} = \left(1 + \frac{\delta\tau}{\tau}\right)^m \quad (12)$$

364 If our hypotheses are correct, a linear relationship is expected between  $\log(1 + \frac{\delta U_s}{U_s})$  and  $\log(1 +$   
 365  $\frac{\delta\tau}{\tau})$  with a slope  $m$ .

366 Surface velocity is the sum, in various proportion, of basal sliding and ice deformation. By  
 367 allowing for differing values of the exponent  $m$ , this model can encompass a range of sliding  
 368 laws, such as those proposed for flow over hard beds with obstacles and without cavitation  
 369 ( $m = 2$ ; ref. [33]); flow over deformable sediment ( $m = 1$ ; ref. [50]); or empirically-derived  
 370 laws from glacier observations ( $m = 3$ ; ref. [34]). Note that  $m \leq 3$  for all of the above sliding  
 371 laws; however, when subglacial pressure is high enough, cavitation (ice-bed decoupling in the  
 372 lee of obstacles) is known to occur. Theoretical work suggests that basal drag is bounded  
 373 independently of sliding velocity [44]. Heuristically we represent this as  $m \gg 1$ : it is not a  
 374 physical model, yet it retains the quality that sliding with cavitation may be more sensitive to  
 375 changes in driving stress than without.

376 Velocity due to ice deformation is represented as (ref. [32], section 8.3):

$$U_d = \frac{2Af}{n+1} \tau^n H \quad (13)$$

377  $A$  is the temperature-dependent creep parameter,  $f$  the valley shape factor and  $n$  the exponent  
 378 of Glen's flow law [51]. For shear stresses taking place in a glacier, a value of  $n = 3$  is generally  
 379 assumed (Ref. [32] section 3.4.4). Assuming all parameters are constant with time, changes in  
 380 driving stress  $\delta\tau$  and thickness  $\delta H$  lead to:

$$1 + \frac{\delta U_d}{U_d} = \left(1 + \frac{\delta\tau}{\tau}\right)^3 \left(1 + \frac{\delta H}{H}\right) \quad (14)$$

381 Considering that  $\frac{\delta H}{H} \approx \frac{\delta\tau}{\tau}$  across all regions investigated (Figure S18,  $R^2=0.97$ ), this can be  
 382 rewritten as:

$$1 + \frac{\delta U_d}{U_d} = \left(1 + \frac{\delta\tau}{\tau}\right)^4 \quad (15)$$

383 also compatible with our observations suggesting  $m = 4$ . This relationship is also compatible  
384 with a contribution, in various proportions, of basal sliding and ice deformation to the surface  
385 velocity (Supplementary section 5).

386

387 In practice, we compute the change in driving stress as follows:

388 (1) We extract ice thickness  $H_{2000}$  and elevation  $S_{2000}$  for year 2000 along centre flow lines  
389 at 50 m spacing (Figure S16a). The centre flow lines have been obtained using the method  
390 proposed by ref. [52] for each glacier of the RGI 5.0. We use ice thickness data provided by ref.  
391 [35]. The data has been validated with ground measurements and the  $1-\sigma$  uncertainty estimated  
392 to 25%. These estimates might not well represent local variations in thickness but provide a  
393 good evaluation of a glacier's average thickness [36]. We use the Digital Elevation Model (DEM)  
394 from the C-band Shuttle Radar Topography Mission (SRTM-C) version 3 acquired in February  
395 2000, available at 1 arc-sec ( $\sim 30$  m) [37].

396 (2) We use elevation change rates, obtained from a series of ASTER-derived DEMs for the  
397 period 2000-2016 [3], to estimate the ice surface  $S_{2016}$  and thickness  $H_{2016}$  for year 2016. To  
398 account for gaps in the data and variability across the glacier, we calculate a mean elevation  
399 change for 50 m altitude bands at each glacier, instead of the centre line value, assuming that  
400 elevation changes are most strongly dependent on mean elevation. We calculate the uncertainty  
401 of the elevation trends for each altitude band using the same methodology as [3].

402 (3) We calculate the surface slope ( $\frac{\partial S_{2000}}{\partial x}$  and  $\frac{\partial S_{2016}}{\partial x}$ ) for both periods using a second order  
403 central difference scheme.

404 (4) We calculate the driving stress along the flow line using Equation 10 and apply a  
405 Gaussian filter of standard deviation  $l = 2H$  to account for the longitudinal coupling of the  
406 stress [53] (Figure S16b).

407 (5) We calculate a relative change in thickness ( $\frac{\delta H}{H} = \frac{H_{2016} - H_{2000}}{H_{2000}}$ ), driving stress ( $\frac{\delta \tau}{\tau} =$   
408  $\frac{\tau_{2016} - \tau_{2000}}{\tau_{2000}}$ ) and associated uncertainties for each point along the flow lines.

409 For comparison with the calculated driving stress, the trend in velocity anomalies is ex-  
410 tracted along the centre flow lines at 50 m spacing. Points with uncertainty in the input  
411 parameters  $\frac{\delta U_s}{U_s}$ ,  $\frac{\delta \tau}{\tau}$  and  $\frac{\delta H}{H}$  larger than 30% are excluded. Finally, a median value of all points  
412 within the RoI is calculated for each region for both the calculated driving stress and the ob-  
413 servations.

414

415 *Uncertainty.* We use a Monte-Carlo method to estimate the uncertainty due to the input  
 416 parameters. We randomly draw (see below)  $H$  and  $\delta H$  to generate an ice surface and thickness.  
 417 We calculate a thickness change and shear stress change profile and repeat the operation 200  
 418 times. We then compute the 68% confidence interval of the distribution in each point.  $H$  has an  
 419 uncertainty of approximately 25% and is positive. Therefore,  $H$  is multiplied by a factor drawn  
 420 from a log-normal distribution with mean 1 and standard-deviation 0.25. The random factor  
 421 is drawn for each glacier individually, to account for the fact that errors in thickness are likely  
 422 correlated for a single glacier due to the way ice thickness is modelled.  $\delta H$  is drawn from a  
 423 Gaussian distribution whose mean and standard deviation are estimated from the distribution  
 424 of values in the elevation band considered. As the  $\delta H$  values are drawn independently at each  
 425 point, this can create step changes in the glacier profile, but, the smoothing used to account  
 426 for the longitudinal stress coupling re-establishes a spatial correlation between neighbouring  
 427 points. We calculate the regional uncertainty from each point's uncertainty  $\sigma_i$  as:

$$\sigma_{reg} = \frac{\sqrt{\sum \sigma_i^2}}{N_{eff}} \quad (16)$$

428 where  $N_{eff}$  is the number of independent points, calculated as the total number of points  
 429 divided by 40. Here, we consider an average correlation length of 2 km as dictated by the  
 430 smoothing (or an average thickness of 250 m), thus 40 points at 50 m spacing.

## 431 Data availability

432 The mean and annual velocity fields will be made publicly available in early 2019 as part of  
 433 the NASA MEaSUREs - ITS\_LIVE project and will be distributed through the National Snow  
 434 and Ice Data centre. Data can be made available immediately through request to the authors.

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## 597 Acknowledgements

598 We thank Matthias Huss for providing the thickness and centre flow line data, Julie Gardelle,  
599 Tobias Bolch, Melanie Rankl and Heidi Sevestre for providing data from their surge-inventory  
600 as well as glacier and basin outlines. We thank Ann Rowan and two anonymous reviewers for  
601 comments and suggestions that greatly improved the quality of the paper. Initial research was  
602 conducted during AD graduate program, with a doctoral fellowship from the Centre National  
603 d’Etude Spatial (CNES) and from the Savoie region. NG and AD were supported by fund-  
604 ing from the European Space Agency Dragon 3 programme. EB acknowledges support from

605 the French Space Agency (CNES). AG and AD were supported by funding from the NASA  
606 Cryosphere and MEaSURES Programs and research was conducted at the Jet Propulsion Lab-  
607 oratory, California Institute of Technology, under contract with the National Aeronautics and  
608 Space Administration.

## 609 **Author contributions statement**

610 A.D., N.G. and A.G. designed the study. A.G. generated the velocity fields. A.D. conducted  
611 the analysis with inputs from A.G and N.G., A.D. developed the model with inputs from D.G.  
612 and P.N. F.B. provided the elevation change data. All authors interpreted the results. A.D.  
613 led the writing of the paper and all co-authors contributed to it.

## 614 **Additional information**

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## 616 **Competing financial interest**

617 The authors declare no competing financial interests.

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