Recent slowdown and thinning of debris-covered glaciers in south-eastern Tibet

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Abstract

Recent large-scale remote sensing studies have shown that glacier mass loss in south-eastern Tibet, specifically in the eastern Nyainqêntanglha Range exceeds the average in High Asia. However, detailed studies at individual glaciers are scarce and the drivers behind the observed rapid changes are poorly constrained to date. Employing feature tracking techniques on TerraSAR-X data for the periods 2008/2009, 2012/2013 and 2013/2014 we found measurable surface velocities through to the glacier terminus positions of five debris-covered glacier tongues. This is contrary to debris-covered glaciers in other parts of High Asia, where stagnant glacier tongues are common. Our feature tracking results for the 2013/2014 period suggest an average deceleration of 51% when compared with published Landsat velocities for the period 1999/2003. Further, we estimated surface elevation changes for the five glaciers from recently released one arc second resolution elevation data obtained during the Shuttle Radar Topography Mission in 2000 and an interferometrical derived TanDEM-X elevation model for the year 2014. With an average rate of $-0.83 \pm 0.57 \text{ m a}^{-1}$ we confirm strong surface lowering in the region, despite the widely discussed insulation effect of debris cover. Beside the influence of thermokarst processes and delayed response times of debris-covered glaciers, we highlight that abundant monsoonal summer rainfall might contribute significantly to the pronounced negative mass balances

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in the study region.

Keywords: glacier velocities, glacier elevation changes, debris-covered glaciers, eastern Nyainqêntanglha Range

1. Introduction

Recent large-scale remote sensing studies based on satellite laser altimetry data and differencing of digital elevation models (DEMs) have shown a heterogeneous pattern of glacier mass changes in High Asia (Gardelle et al., 2013; Neckel et al., 2014; Kääb et al., 2015). While balanced or slightly positive mass balances are evident for parts of the Pamirs and the Karakorum in recent years, pronounced negative glacier mass balances were found in southeastern Tibet, specifically in the eastern Nyainqêntanglha Range. Details about these glaciers' specific response to climate change are, however, scarce, because few local glaciological mass balance measurements exist and multi-annual records are lacking. Furthermore, meteorological station records are rare and can not directly be transferred to the glacierized high mountain regions, as most stations are located in inhabited river valleys. Even if glaciers are located in the same climatic environment, local parameters such as catchment aspect, topography and debris cover can have a significant impact on how a specific glacier responds to climate change (Kääb, 2005; Scherler et al., 2011).

The combination of high mountain topography and abundant moisture facilitates $\sim 8000 \text{ km}^2$ of glaciated area in the eastern Nyainqêntanglha Range, including some of the largest glaciers in High Asia (Shi et al., 2010; Hochreuther et al., 2015; Loibl et al., 2015). This study focuses on five debris-covered valley glaciers in the central-northern part of the mountain range, specifically in the upper reaches of Lequ Valley and Bodui Valley (Figure 1). Glaciers in the study area are reported to be of temperate type and are influenced by the South Asian monsoon circulation (Shih et al., 1980; Huang, 1990). These glaciers receive most precipitation in summer and/or spring, which falls as snow in their high altitude accumulation areas, resulting in the coincidence of maximum accumulation and ablation phases (Maussion et al., 2014; Loibl et al., 2014). The lower portions of many temperate glaciers in this region are covered by supraglacial debris (Wei et al., 2010). Debris-covered glaciers are known to react differently to warming when compared to clean-ice glaciers, as the debris cover insulates the glacier surface provided a sufficient thickness

of a few centimeters is reached (Östrem, 1959; Reznichenko et al., 2010). Due to this insulating effect debris-covered glaciers can extend to lower elevations than clean-ice glaciers would in a similar geographic environment (Rowan et al., 2015). However, this insulating effect is well known on local scales, but it remains unclear whether it can be applied to entire glacier tongues (Kääb et al., 2012).

Observations of glacier velocities and surface elevation changes are important parameters to better understand the current state of glaciers and their interaction with changes in regional climate conditions (Cuffey and Paterson, 2010; Paul et al., 2015; Thakuri et al., 2016). In this study, we applied a feature tracking approach based on TerraSAR-X data to derive glacier surface velocities for the periods 2008/2009, 2012/2013 and 2013/2014. The resulting velocity fields were compared to optical remote sensing-based estimates for the period 1999/2003 as published by Dehecq et al. (2015). The velocity fields and their interannual variability were analyzed to investigate glacier dynamics in this key region where no *in-situ* measurements are available. Further we discuss the differences to other debris-covered glaciers in the Himalayas and on the Tibetan Plateau. Additionally, we employed data from the TerraSAR-X add-on for Digital Elevation Measurement (TanDEM-X) mission to calculate recent glacier surface elevations. In order to estimate glacier surface elevation changes between 2014 and 2000 the TanDEM-X elevations were compared to digital elevation data from the Shuttle Radar Topography Mission (SRTM). Here, preference was given to the recently released one arc second version of the dataset. By conducting an integrated analysis of measured glacier velocities and surface elevation changes we investigated spatial and temporal properties of glacier dynamics against the background of the monsoonal climate setting in the study area.

2. Data and methods

2.1. Glacier surface velocities

Surface velocities between the periods 2008/2009, 2012/2013 and 2013/2014 were derived from intensity/feature tracking on repeat-pass TerraSAR-X satellite imagery (see Rankl et al., 2014, and references therein, for details). Based on the image intensity, this technique tracks surface features and, if coherence is retained, the speckle pattern on a pair of co-registered, single-look complex (SLC) images from two different acquisition dates. The tracking algorithm uses the maximum of cross correlation in a pre-defined moving

Figure 1: Regional overview of the study area in the eastern Nyainqêntanglha Range, south-eastern Tibet. Location of Figure 2 and the five investigated glaciers are indicated. Background: Landsat-7 ETM+, bands 5-4-3-pan, acquired on 23 September 1999.

search window. For this, a window size and a step size between patches need to be defined. After testing different search window sizes, we chose a final window size of both 256 pixels in range and azimuth with a step size between patches of 25 pixels in range and azimuth. This setting was found to be most robust considering the long time interval between data acquisitions. Table 1 lists the TerraSAR-X scenes which were used for the synthetic aperture radar (SAR) offset tracking. Surface velocities for the period 1999/2003 are based on Landsat estimates following the method of Dehecq et al. (2015). The velocity field was obtained from a stack of velocities derived from feature tracking on Landsat-7 band 8 image pairs separated by approximately one year. For the final stack, all available Landsat-7 acquisitions over the period 1999/2003 were used (Dehecq et al., 2015, A. Dehecq, personal communication, March 2016).

2.2. Surface elevation changes

Surface elevation changes (Δh) were calculated between data from the Shuttle Radar Topography Mission (SRTM) acquired in 2000 (Rabus et al., 2003; Farr et al., 2007) and the 2014 TanDEM-X acquisition, which was also employed for the 2013/2014 velocity field (Table 1). Here, we make use of the SRTM C-band DEM with a grid spacing of one arc second, which was recently released by the U.S. Geological Survey (USGS). The TanDEM-X DEM was derived with a similar processing scheme as described by Neckel et al. (2013). In order to minimize unwrapping errors in the processing of the TanDEM-X DEM, phase unwrapping was performed on a differential interferogram. The latter was formed by subtracting the simulated phase of the SRTM DEM from the TanDEM-X interferogram. A remaining phase ramp was removed by applying a 2D quadratic phase model on the unwrapped differential interferogram. The final TanDEM-X DEM was generated by adding back the differential heights to the SRTM DEM. Horizontal offsets between the SRTM DEM and the TanDEM-X DEM were eliminated by using the precise orbit information of the TanDEM-X data and a refinement of the lookup table, which holds the relation between the map coordinates of the SRTM DEM and the matching TanDEM-X master scene available in SAR geometry. The refinement of the lookup table is based on a horizontal co-registration between a simulated SAR image of the SRTM DEM and the TanDEM-X master scene using an intensity cross-correlation algorithm (e.g. Rankl and Braun, 2016).

Our final elevation changes between the SRTM DEM and the TanDEM-X

DEM were calculated following the approach of Gardelle et al. (2013). This approach is based on the assumption that elevation changes should be similar at a given altitude. Therefore Δh values of each glacier were grouped into 100 m altitude bands. Following Gardelle et al. (2013) outliers were removed, if they deviated by more than three standard deviations from the Δh mean of the respective altitude band. In order to calculate average elevation changes for the entire glacier area “no data” values were filled with the cleaned Δh mean of the appropriate altitude band.

Glacier DEMs derived by SAR interferometry are potentially biased by penetration effects of the radar waves into snow and ice and hence, might reflect some lower surface than the actual glacier surface (Rignot et al., 2001; Gardelle et al., 2012). This holds true especially in the accumulation area as largest penetration depths can be assumed for snow and firn compared to smaller values for ice (Kääb et al., 2015). Without additional information of the surface characteristics at the time of data acquisition, this bias hardly can be quantified and as a consequence, is widely discussed in the literature (Berthier et al., 2006, 2007; Kääb et al., 2012; Gardelle et al., 2013; Holzer et al., 2015). Here we rely on published values from Kääb et al. (2015) who extrapolated Δh trends derived from Ice Cloud and Elevation Satellite (ICE-Sat) measurements to 2000 for several sub-regions in High Asia. Following the estimate of Kääb et al. (2015) for the eastern Nyainqêntanglha Range, we applied a penetration depth correction of 8.5 ± 1.5 m in the accumulation area. We considered no radar penetration in the ablation area as the glaciers are extensively debris covered. We did not account for the X-band penetration of TanDEM-X, but assume that it is much smaller when compared to the C-band penetration of SRTM (Gardelle et al., 2012).

2.3. Glacier areas

Glacier masks are based on the Landsat estimates from Loibl et al. (2014). Outlines from the original mapping were extended to cover all potentially debris-covered areas, omitting the visual interpretation of stagnant ice areas. This modification was necessary to investigate glaciers as a whole and to evaluate the optical interpretation of stagnant ice areas as described in detail in Loibl and Lehmkuhl (2015). The Landsat-7 ETM+ acquisition from 23 September 1999 remains the only Landsat scene fulfilling the required criteria for appropriate glacier mapping in the frequently cloud and snow covered study area. In order to separate the accumulation and ablation areas of the

Table 1: Remote sensing datasets used for glacier masks, estimates of the Transient snowline (TSL), Δh calculations and feature tracking.

Acquisition date	Mission	Relative orbit	Application
1999-09-23	Landsat-7 ETM+	135	Glacier masks and TSL
2000-02-11	SRTM	-	Δh
2008-09-27	TerraSAR-X	135	Feature tracking
2009-09-03	TerraSAR-X	135	Feature tracking
2012-07-05	TerraSAR-X	158	Feature tracking
2013-07-03	TerraSAR-X	158	Feature tracking
2013-06-26	TerraSAR-X	59	Feature tracking
2014-06-02	TanDEM-X	59	Feature tracking and Δh

143 five investigated glaciers, we manually delineated the late summer Transient
 144 snowline (TSL) for each glacier from the Landsat scene. Debris-covered areas
 145 of the five studied glaciers were mapped manually (cf. Figures S2-S5). All
 146 area estimates were translated from planimetric areas into surface areas by
 147 taking into account the local surface slope obtained from the SRTM DEM
 148 (Lopez and Berry, 2002).

149 *2.4. Data quality and uncertainties*

150 As the time interval between the TerraSAR-X acquisitions is between 341
 151 and 363 days, no speckle correlation is preserved on clean ice between the
 152 data acquisitions due to loss of signal coherence. Similar to results from op-
 153 tical feature tracking, the offset tracking analysis therefore relies on surface
 154 structures such as crevasses and debris cover. Any surface motion over struc-
 155 tureless, homogeneous areas may be undetectable. Accordingly, the method
 156 works particularly well for tongues of large debris-covered valley glaciers as
 157 indicated by consistent flow directions in the derived velocity fields (Figures
 158 S2-S5). Accumulation areas and steep passages represent major challenges to
 159 the method as indicated by arbitrary flow directions in these regions. There-
 160 fore we applied an additional directional filter with a threshold of $\pm 20^\circ$ be-
 161 tween the TerraSAR-X derived surface velocities and the 1999/2003 Landsat
 162 velocity field (e.g. Scherler et al., 2008). Additionally, the rugged topography
 163 poses challenges to radar data acquisition, i.e. shadowing and layover effects,
 164 resulting in striped artifacts and data gaps. The latter also applies for the
 165 interferometrically derived TanDEM-X DEM.

166 Considering the limitations outlined above, the debris-covered tongues of five
 167 large valley glaciers showed the most coherent results in terms of surface ve-
 168 locities and elevation changes (Figure 2). Owing their diversity in length,

169 altitude, spatial orientation and degree of debris cover, they cover a broad
 170 spectrum of glaciers in the wider region (Figure 1).
 171 In order to quantify errors of the derived surface velocity fields, we assume
 172 that no motion should be observable in *off-glacier* regions, i.e. on bedrock.
 173 Displacement values collected over these areas were considered erroneous
 174 (Copland et al., 2009; Quincey et al., 2011; Rankl et al., 2014; Dehecq et al.,
 175 2015; Seehaus et al., 2015). *Off-glacier* values were defined as all data values
 176 located in a 1 km buffer outside the glacierized area with a surface slope
 177 between 0° and 30°. The latter was chosen as it represents the approximate
 178 range of the glacier surface slope in the study area. From these data samples,
 179 errors in surface velocities were calculated by (Mcnabb et al., 2012; Seehaus
 180 et al., 2015)

$$181 \quad e_{vel} = 365 \frac{C \Delta x}{z \Delta t} + \sigma \quad (1)$$

182 where C is an uncertainty estimate of the offset tracking algorithm, Δx is the
 183 SLC image resolution, z is the applied SLC oversampling factor and Δt is
 184 the time difference between the two data takes employed in the offset track-
 185 ing. For the 2008/2009, 2012/2013 and 2013/2014 velocity fields, Equation
 186 1 resulted in $\pm 1.07 \text{ m a}^{-1}$, $\pm 1.45 \text{ m a}^{-1}$ and $\pm 5.53 \text{ m a}^{-1}$, respectively.
 187 The error calculation of the Δh grid is based on *off-glacier* values meeting
 188 the same requirements as for the velocity error estimate, i.e. all Δh values
 189 located in a 1 km buffer outside the glacierized area with a surface slope
 190 between 0° and 30° are considered. Here, we calculated the Normalized Me-
 191 dian Absolute Deviation (NMAD) of the respective *off-glacier* data samples
 192 (e.g. Höhle and Höhle, 2009). The resulting mean and NMAD are 4.79 m
 193 and 8.07 m respectively. The final Δh grid was corrected for the mean value.
 194 For the overall error of the average glacier elevation changes the NMAD and
 195 the uncertainty of the radar penetration depth were added in quadrature.
 196 For the average Δh estimates a certain error due to the filling of “no data”
 197 values is introduced, which is difficult to quantify. Here, we argue that our
 198 results are within the errors bars when compared to results without filling
 199 data gaps in the Δh grid.

200

201 3. Results

202 The velocity data of the five investigated debris-covered glaciers indicate
 203 that their terminus regions are still active. Surface velocities are lowest at the

Figure 2: TerraSAR-X feature tracking results between 2013-06-26 and 2014-07-02, profiles are labeled according to Table 2 and Figure 3 (a). Surface elevation changes (Δh) between 2000 and 2014 are based on SRTM and TanDEM-X data respectively (b).

glacier fronts but increase with elevation (Figures 2a and 3). For Zepu Glacier an almost linear increase of surface motion is observed, until a maximum is reached approximately 5 km from the terminus. Then surface velocities remain high up to the end of the valid velocity measurements. For the 2013/2014 period maximum velocities between 60 m a^{-1} and $\sim 70 \text{ m a}^{-1}$ are found in this region (Figure 3). High surface velocities of $\sim 80 \text{ m a}^{-1}$ are also evident close to the junction of a tributary joining the glacier tongue of Unknown Glacier I from the north (Figure 2a).

To assess multi-annual trends in glacier velocities, longitudinal profiles of yearly average velocities were obtained along the centerlines of the five investigated glaciers (Figure 3). All five glaciers experienced a decrease in

215 velocity during the time of investigation. Deceleration of surface velocities
 216 affects each glacier tongue, but glaciers differ in the amount of slowing. Along
 217 the profiles shown in Figure 3 the highest rate of deceleration between the
 218 1999/2003 period and the 2013/2014 period is observed for Unknown Glacier
 219 I. Its average speed decreased by 63% from 31.79 m a^{-1} during the 1999/2003
 220 period to 11.61 m a^{-1} during the 2013/2014 period (Table S1). Lowest rates
 221 of deceleration are found at Unknown Glacier II. While its average speed
 222 was 23.79 m a^{-1} during the 1999/2003 period, it slowed down by 43% to
 223 13.57 m a^{-1} during the 2013/2014 period (Table S1). On average, the five
 224 investigated glacier tongues slowed down by 51% between the 1999/2003 pe-
 225 riod and the 2013/2014 period based on the valid measurements along the
 226 centerline profiles shown in Figure 3.

227 For Unknown Glacier I and Maguolong Glacier the change in surface ve-
 228 locity between the 2008/2009 period and the 2012/2013 period (4 years
 229 time difference) is only minor, but is followed by a slowdown between the
 230 2012/2013 period and the 2013/2014 period (1 year time difference) with a
 231 similar magnitude as the velocity change between the 1999/2003 period and
 232 the 2008/2009 period (~ 7 years time difference). By contrast, Zepu Glacier
 233 exhibits more consistent deceleration between observations. However, when
 234 keeping in mind that the time intervals between the observational periods are
 235 not equally spaced, one can assume a more pronounced deceleration in more
 236 recent years. Further, it should be noted, that possible variations in glacier
 237 velocity between 1999 and 2003 can not be assessed, as the Landsat-derived
 238 velocity field is a composite of all available data acquired before the failure
 239 of the scanline corrector of Landsat-7 in order to increase the accuracy of the
 240 flow field (A. Dehecq, personal communication, March 2016).

241 Differential analysis of SRTM and TanDEM-X DEMs covering approximately
 242 the same time period as the velocity estimates (2000-2014), revealed associ-
 243 ated changes in surface elevation. During this time period, the five investi-
 244 gated glaciers show substantial surface lowering with an area-weighted mean
 245 of $-0.83 \pm 0.57 \text{ m a}^{-1}$. With an average rate of $-1.40 \pm 0.57 \text{ m a}^{-1}$ Guanxing
 246 Glacier exhibits the most negative elevation changes of the five investigated
 247 glaciers. Conversely, Unknown Glacier I shows the least negative elevation
 248 changes of $-0.50 \pm 0.57 \text{ m a}^{-1}$ on average (Table 2). Most pronounced surface
 249 lowering is apparent at the tongue regions of the investigated glaciers. While
 250 Zepu Glacier, Unknown Glacier I and Unknown Glacier II show most neg-
 251 ative Δh values 2.5 km to 5 km from the terminus (Figures 3, S2 and S4),
 252 Guanxing Glacier and Maguolong Glacier show strongest surface lowering

near their terminus positions (Figures S3 and S5 respectively). For Guanxing Glacier and Unknown Glacier I strong surface lowering is also evident at the junctions of major tributaries. Even though several accumulation areas yield positive elevation changes, they cannot compensate for the strong mass loss in the ablation areas.

Comparison between the 1999 glacier outlines and the geocoded TanDEM-X master scene from 2014 show only a minor retreat of the glacier terminus positions during the time period of observation. An exception is found for Guanxing Glacier, where a proglacial lake is forming at the glacier terminus (Figure 2b). With a total retreat of ~ 500 m between 1999 and 2014 Guanxing Glacier exhibits the largest terminus retreat of the five investigated glaciers.

Table 2: Glaciers investigated in detail. Glacier names are listed next to the estimated altitude of the Transient snowline (TSL) on 23 September 1999, the corresponding surface area of the accumulation, ablation and glacier area as well as the percentage of debris cover. Further, the mean surface elevation changes ($\overline{\Delta h}$) between 2000 and 2014 are shown.

Glacier	TSL elevation 1999 (m)	Accumulation area (km ²)	Ablation area (km ²)	Glacier area (km ²)	Debris- covered area (%)	$\overline{\Delta h}$ (m a ⁻¹)
Zepu	4828	52.86	35.42	88.28	14	-0.58 $\pm$ 0.57
Unknown I	4900	48.58	20.54	69.12	11	-0.50 $\pm$ 0.57
Unknown II	4993	6.82	5.62	12.44	27	-0.67 $\pm$ 0.57
Maguolong	5072	37.60	24.15	61.75	15	-1.19 $\pm$ 0.57
Guanxing	5069	29.92	13.96	43.88	9	-1.40 $\pm$ 0.57

4. Discussion

Throughout the periods of observation, the five debris-covered glacier tongues investigated in this study show evidence of surface motion through to their terminus regions. This finding is contrary to studies in the Himalayas, where many glacier tongues were found to be stagnant (Quincey et al., 2009; Scherler et al., 2011). The fact that surface velocities are measurable through to the glacier terminus suggests extensive accumulation in the higher regions of the investigated glaciers ensuring sufficient ice flux to the glacier tongues. This is in accordance with positive Δh values in the highest portions of the accumulation areas, being particularly evident at the large high-altitude plateau of Unknown Glacier I (Figure S4). Even though uncertainties of radar penetration into snow, firn and ice make an in-depth analysis of these effects difficult, the spatial distribution of positive Δh values is consistent with expectations regarding regional climate observations.

Figure 3: Multi-annual surface velocities and 2000/2014 surface elevations along the centerlines of the five investigated debris-covered glacier tongues (see Figure 2 for the locations of the profiles and Table 2 for details on the glaciers). Landsat-derived surface velocities for the 1999/2003 period are provided by Dehecq et al. (2015). Note, velocities are equally scaled.

Specifically, monsoonal precipitation is very intense (Böhner, 2006; Maus-ion et al., 2014) and tends to increase in the study region (You et al., 2012; Kim and Park, 2016). At the lower portions of the glaciers monsoonal precipitation falls as rain during summer, potentially enhancing melt in these areas. This is in agreement with findings from nearby Parlung No. 4 Glacier (29.24° N, 96.92° E), where most rapid melting occurs during the monsoon core season. In this time period the release of latent heat and the energy input from incoming longwave radiation have been shown to have a strong effect on ablation (Yang et al., 2011).

The five debris-covered glacier tongues investigated in this study show a substantial slowdown between the 1999/2003 period and the 2013/2014 period, which is in general agreement with recent glacier deceleration in other parts of High Asia (e.g. Thakuri et al., 2016) and other mountain ranges elsewhere (e.g. Span and Kuhn, 2003; Berthier and Vincent, 2012; Heid and Kääb, 2012). Glacier slowdown was also reported for the neighboring Hailuogou Glacier (29.59° N, 101.95° E), where the average annual velocity in the ablation area reduced by 24% between 1981 and 2008 (Zhang et al., 2010). The deceleration of the debris-covered Hailuogou Glacier is in line with our ob-

296 servations, which show an average annual deceleration in the ablation area
 297 between 18% and 36% between the 1999/2003 period and the 2008/2009
 298 period (Table S1). Similar to our results, Hailuoguo Glacier shows motion
 299 through to its terminus region (Zhang et al., 2010).
 300 The five glaciers investigated in detail show negative average elevation changes
 301 between 2000 and 2014. This is in general agreement with large-scale esti-
 302 mates based on ICESat laser altimetry and DEM differencing (Gardelle et al.,
 303 2013; Neckel et al., 2014; Kääb et al., 2015). Largest differences are evident
 304 when compared to the results of Gardelle et al. (2013) based on DEM differ-
 305 encing. Most of these deviations can be explained by the applied estimates
 306 when correcting for C-Band penetration depth. While Gardelle et al. (2013)
 307 employed a penetration depth correction of 1.7 ± 1.5 m, we used a value of
 308 8.5 ± 1.5 m following Kääb et al. (2015). Further, the time span between ob-
 309 servations is different between both studies. However, our observations agree
 310 within their error bars with the ICESat based studies of Neckel et al. (2014)
 311 and Kääb et al. (2015) and when applying a C-band penetration depth cor-
 312 rection of 1.7 ± 1.5 m also with Gardelle et al. (2013).
 313 Glaciers with negative mass balances are known to decelerate as less mass
 314 is transported down the glaciers (Heid and Kääb, 2012). This holds true if
 315 the accumulation volume is reduced at the glaciers' higher portions. Several
 316 studies observed an increase in air temperatures on the Tibetan Plateau dur-
 317 ing the last decades (e.g. Liu and Chen, 2000; Yang et al., 2014; Duan and
 318 Xiao, 2015). This might reduce the extent of accumulation areas, resulting
 319 in less snow accumulation and therefore in a decrease of ice flux and hence,
 320 surface velocities (Loibl et al., 2014).
 321 Even though all investigated glaciers show negative average elevation changes,
 322 except for Guanxing Glacier frontal positions were relatively stable during
 323 the period of observation. This indicates that these glaciers are losing mass
 324 with only minor terminus retreat, a phenomenon which is well known for
 325 debris-covered glaciers (e.g. Benn et al., 2012; Rowan et al., 2015). Our data
 326 show that a previous study of Loibl et al. (2014), which applied optical re-
 327 mote sensing to map glacier extents in the same region, underestimates the
 328 positions of glacier fronts. This is attributed to the limitations of optical
 329 identification of debris-covered stagnant ice areas. Specifically, parts of the
 330 tongues, which were previously thought to be decoupled from the main glacier
 331 trunk, were still moving according to the velocity data presented here.
 332 Detailed investigations of glacier elevation changes revealed strong down-
 333 wasting for the entire tongues of all investigated glaciers. For Zepu Glacier,

334 Unknown Glacier I and Unknown Glacier II maximum elevation changes oc-
 335 curred not directly at the glacier terminus but rather 2.5 km to 5 km up
 336 glacier (Figures 3, S2 and S4). Such a pattern is known for debris-covered
 337 glacier tongues as debris cover thickens down glacier and therefore enhances
 338 its insulating effect (Nakawo et al., 1999). This can cause a reduction of the
 339 glacier gradient, resulting in a reduction of driving stresses and ice velocities
 340 (Quincey et al., 2009; Benn et al., 2012; Rowan et al., 2015). By contrast,
 341 Maguolong Glacier shows most negative elevation changes directly at its ter-
 342 minus despite its debris-covered tongue (Figure S5). Such a pattern is most
 343 common for clean-ice glaciers and is also found at the less debris-covered
 344 Guanxing Glacier (Figure S3). Guanxing Glacier also shows the most nega-
 345 tive glacier-averaged elevation changes (Table 2), which can be attributed to
 346 two reasons: (1) there is less insulation at the glacier terminus due to miss-
 347 ing debris cover and (2) the terminus is directly affected by a proglacial lake
 348 (Figure 2b). Both points might also explain the strongest terminus retreat
 349 in our study area during the time of observation. Least negative average
 350 elevation changes are evident for Unknown Glacier I (Table 2), which can be
 351 attributed to the fact that the glacier is fed by a large high-altitude accumu-
 352 lation area (Figure S4).

353 Overall, our results show substantial downwasting for entire glacier tongues
 354 even if they are covered by thick debris. This indicates that for the eastern
 355 Nyainqêntanglha Range the frequently observed insulation function of debris
 356 cover takes effect on a local scale, but is impaired on a larger scale of entire
 357 glacier tongues. In the following, three possible scenarios are briefly discussed
 358 which attempt to explain the observed pattern: (1) prominent average sur-
 359 face lowering of debris-covered glaciers might be influenced by thermokarst
 360 processes (Kääb et al., 2012). In several studies it has been shown that
 361 melting of ice cliffs can significantly contribute to the overall ablation of
 362 debris-covered glaciers (Reid and Brock, 2014; Buri et al., 2016). Ice cliffs
 363 are present on the investigated glaciers and their melting might contribute
 364 to the observed enhanced glacier-averaged surface lowering (Figure S1). (2)
 365 A recent modeling study suggests that glacier-averaged surface lowering is
 366 smaller for debris-covered glaciers than for clean-ice glaciers at an initial
 367 stage of warming (Banerjee, 2017). However, at a more advanced stage of
 368 glacier evolution melt rates of debris-covered glaciers equal or even overtake
 369 that of clean-ice glaciers due to longer response times. This encourages the
 370 assumption, that the investigated glaciers show an advanced stage of neg-
 371 ative mass balance. (3) Abundant monsoonal summer rainfall may bypass

the debris barrier and carries heat to the underlying ice bodies. Laboratory studies suggest that the thickness and permeability of debris cover have a significant effect on the amount of rain-enhanced melt underneath the debris layer (Reznichenko et al., 2010). While the effect beneath medium sand is equal to that of bare ice, thick rock-avalanche debris are able to reduce rainfall enhanced ablation. As heat flux due to precipitation tends to be small for many glaciers, it is often neglected in modeling studies. However, specifically for monsoonal-influenced glaciers, this could be an important heat flux (Reid and Brock, 2010).

While point (1) and (2) can also explain the observed anomaly of debris-covered glaciers in other parts of High Asia (e.g. Kääb et al., 2012; Banerjee, 2017), point (3) is more site-specific, as it requires strong monsoonal rainfall on the lower portions of glaciers. Which of the above mechanisms is the main source for the strong average surface lowering observed in the study region cannot fully be answered by the presented remote sensing dataset requiring further field measurements. However, here we highlight the abundant monsoonal summer rainfall, which may contribute significantly to the regions' high mass losses when compared to predominantly Westerly-influenced regions on the western Tibetan Plateau and in the Pamirs (Yao et al., 2012).

5. Conclusions

In this study, new datasets of multi temporal glacier surface velocities and glacier elevation changes were presented for five debris-covered glaciers in the eastern Nyainqêntanglha Range, south-eastern Tibet. The investigated glaciers experienced a significant deceleration between 1999/2003 and 2013/2014, but showed measurable surface velocities through to their terminus positions during all time periods of observation. Compared to observations from other regions in High Asia, where most debris-covered glacier tongues show large parts of stagnant ice, this is exceptional. Here, we argue that this is due to intense monsoonal accumulation in the higher glacier regions. By detailed analyses of glacier elevation changes between 2000 and 2014 we observed a reduction of the surface gradient of three glaciers, which can partly be attributed to the spatial distribution of debris cover altering the pattern of glacier melt. Nevertheless we confirm pronounced negative glacier elevation changes in the study region of $-0.83 \pm 0.57 \text{ m a}^{-1}$ on average. This could be attributed to thermokarst processes or a delayed response time of debris-covered glaciers as observed in other parts of High Asia. Further-

408 more, heavy monsoonal summer rainfall may additionally enhance glacier
409 melt in the study region. In order to investigate the contribution of those
410 forcing mechanisms to local glacier mass balances further field measurements
411 are required.

412 **Author contributions**

413 M. Rankl derived the velocity fields by means of TerraSAR-X feature
414 tracking. N. Neckel calculated the surface elevation changes and did the
415 post processing of the TerraSAR-X velocity fields. D. Loibl conducted GIS
416 analyses. N. Neckel and D. Loibl wrote the manuscript.

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