Geomorphic consequences of rapid deglaciation at Pasterze Glacier, Hohe Tauern Range, Austria, between 2010 and 2013 based on repeated terrestrial laser scanning data

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Abstract

Since the end of the Little Ice Age around 1850 CE glaciers in the Alps have been receding dramatically. This study aimed to quantify and characterize the geomorphic and landform changes of a 0.9 km² large proglacial area at the largest glacier in Austria (Pasterze Glacier, Austria, N 47°04′, E 12°44′). Point clouds from multiple terrestrial laserscanning (TLS) and different image data were used to quantify surface elevation changes and distinguish different types of erosional and depositional landforms during the period 2010–2013. Results indicate that the study area is characterized by a total volume loss of 1,309,000 m³. Excluding the area which was deglaciated, the volume loss equals 275,000 m³ in the period 2010–13. The decrease is related to sediment transfer out of study area and due to sediment-buried glacier ice which is slowly melting. The landform classification reveals that drift mantled slopes are most frequent (20.9% of the study area in 2013) next to ice contact terrace landforms (19.7%).

1. Introduction

Since the last period of extensive glaciation during the Little Ice Age (LIA, around 1850 CE) most glaciers in the European Alps have been retreating continuously with only short periods of glacier readvances or stagnations (Haeberli et al., 2007). In Austria, glaciers covered about 940 km² during the LIA, about 565 km² in the late 1960s, about 472 km² in the late 1990s, and 415 km² around 2007 (Fischer et al., 2015). Large alpine valley glaciers still terminate at lower elevations, a result of the long response time of such glacier systems to climate warming. However, at lower elevation climatic conditions are generally unfavorable for glacier persistence due to warmer air temperatures. This leads to a loss of accumulation area and high rates of ice mass loss causing rapid glacier recession and surface lowering.

The deglaciated area is characterized by high geomorphic activity explained by the paraglacial landscape concept (e.g. Ballantyne, 2002). Glacially eroded and transported material is stored in unconsolidated, mostly unstable and often transient landforms. These landforms are created in several generations, for example nested terrace systems as a result of steady glacier surface lowering. Unvegetated drift-mantled slopes are very sensitive to erosion with gully formation and related debris cone deposition further below (Ballantyne and Benn, 1994). Glacial rivers redistribute large amounts of glacial sediments in the proglacial sandur area and transport this material out of proglacial area (Ballantyne, 2002).

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https://doi.org/10.1016/j.geomorph.2018.02.003
0169-555X/© 2017 Published by Elsevier B.V.
Monitoring such rapid changes with remote sensing techniques at a high spatial and temporal resolution is crucial for process quantification and hence understanding. Nowadays a frequently used tool for monitoring proglacial areas is either airborne laserscanning/ALS (e.g. Sailer et al., 2014) or terrestrial laserscanning/TLS (e.g. Gehlhauser et al., 2012; Heckmann et al., 2012). Based on such ALS and TLS data high-resolution digital terrain models (DTM) (≤ 1 m raster cell size) can be produced for further analyses. In this decade TLS has been widely used for monitoring glacial and paraglacial dynamics in the Alps characterizing e.g. sediment cascades and quantifying sediment budgets (e.g. Fischer et al., 2012; Baewert and Morche, 2014). Carrivick et al. (2013) applied contemporary multi-scale determination of geomorphological changes at the proglacial area of the Ödenwinkelkees (Austria). Periglacial applications such as quantifying rock glacier dynamics were carried out by e.g. Bodin and Schoeneich (2008) and Avian et al. (2009). Methodological considerations in the use of repeated laserscanning for geomorphological applications were suggested by Abermann et al. (2010).

This paper aims to contribute to the quantification and characterization of landscape changes in a geomorphologically highly active proglacial area. In doing so we analyzed an area of 0.886 km² in front of Pasterze Glacier, quantifying and discussing the spatial variability and magnitudes of processes along with the evolution of landforms within the area which has been deglaciated since 1980.

2. Pasterze glacier and study area

Pasterze Glacier is the largest glacier in the Eastern Alps covering an area of 17.3 km² in 2009 (Kaufmann et al., 2015). The glacier is located east of the Großglockner mountain (3798 m a.s.l.) with its terminus at N 47°04’, E 12°44’. Regular glaciological monitoring at Pasterze Glacier was initiated as early as 1878 resulting in one of the longest time series of glacier behaviour in the Alps (Wakonigg and Lieb, 1996). Pasterze Glacier reached its last maximum at the end of the LIA in 1852/56 (precise year not known; Nicolussi and Patzelt, 2000a) covering an area of 26.5 km² (Paschinger, 1969). Since then Pasterze Glacier has been retreating consistently apart from a short period of stagnation around 1910–1940 (Wakonigg and Lieb, 1996). During the Holocene the glacier was several times smaller compared to today as indicated by glacially reworked peat and wood findings (Nicolussi and Patzelt, 2000b; Kellerer-Pirklbauer and Drescher-Schneider, 2009).

Bedrock is dominated by prasinite (a type of greenschist) and amphibolite at the south western Grossglockner-ridge and calcareous mica schist at the north-eastern Fuscherkarkopf-ridge (Höck and Pestal, 1994). Hanging glaciers, which were once connected to the tongue of Pasterze Glacier, contribute to the drainage system from the NE-facing Grossglockner ridge. In contrast, the opposite valley side of the Fuscherkarkopf-ridge facing southwest is presently free of glacier ice. The tongue of Pasterze Glacier is characterized by a distinct supraglacial debris-cover (in the SW) which is in contrast to a relatively clean ice part (in the NE, see Fig. 2). The debris mantle causes substantial different ablation behaviour of the two parts (Kellerer-Pirklbauer et al., 2008) which also influences the glacier flow pattern (Kaufmann et al., 2015). The area of the debris cover at the Pasterze Glacier increased substantially during the last four decades, covering about 72% of the glacier tongue in 2009 (Kellerer-Pirklbauer, 2008; Kaufmann et al., 2015).

The selected study area (Fig. 1/C) comprises a part of the proglacial region of the Pasterze Glacier near the present glacier terminus. The spatial extent of the study area is defined by the practical maximum measurement distance of the used TLS sensor LMS Z620, about 1000 m covering some 0.885 km² (Fig. 1C). This area has been deglaciated since c.1980 as known from e.g. color infrared images from 1983 (cf. Fig. 1C). The study area therefore includes different stages of proglacial landscape modification although most of it was deglaciated during the last 35 years.

3. Terrestrial laserscanning and GPS

3.1. Data acquisition

Laserscanning is mainly used to create high resolution DTMs expressed in triangulated irregular networks (TIN) or in raster formats. These data sets are 2.5D representations of the earth's surface, while a genuine point cloud is a 3D representation. The original point clouds are often used for calculating distances between homologue points referred to as movements in multi-temporal data sets using matching algorithms such as “iterative closest points” (ICP: Besl and McKay, 1992), or “least squares matching” (LS3D: Gruen and Akca, 2005).

The quality of point clouds from TLS mainly depends on the position of the sensor to the object, which can cause unfavorable geometrical conditions such as: (i) shadowed areas, (ii) a flat scanning angle, or (iii) large distances to the object leading to large footprints (i.e. large area of the laser profile at the object and/or large ground sampling distance = GSD). As a crucial parameter for the quality of the resulting point cloud, the typical theoretical ranging accuracy of long range laser scanners (such as the RIEGL LMS Z620 used here) is ±10 mm + 1 cm/100 m under test conditions. Wet weather or very bright ambient conditions, the scanning geometry (incidence angle, footprint), and/or poor reflective surfaces usually minimize the maximum range considerably.

For this study TLS measurements were carried out with the sensor RIEGL LMS Z620 (wavelength 1550 nm, theoretical ranging distance up to 2000 m, for more details see RIEGL manual for LMS Z620). The surface of relatively clean ice is not detectable by this sensor due to the inherent wavelength of the scanner LMS Z620 of 1550 nm. A 0.886 km² large area was scanned in 2010, 2012, and 2013. In 2011 a slightly smaller area of 0.748 km² was scanned due to the effects of a different scanning geometry (a manual device for scanner tilting was not available).

By August 2000 a geodetic network consisting of seven stable points had already been established (one scanning position and six reference points). During that time the RIEGL LPM-2k was used particularly for glacier monitoring (Avian et al., 2007). All established reference points are located at a distance of 50 to 150 m to the scanning position. The reference targets used (RIEGL circular reflective targets, diameter 10 cm) are fixed on infrastructure or bedrock with bolts and are re-measured every year for quality control. These targets were used for the exterior orientation in the registration of each TLS measurement showing an accuracy of annual measurements of ±2.2–±2.6 cm (Table 2). The chosen scanning position is located on the roof of a cable car station (Fig. 2) in the tourist area Franz-Josefs-Höhe (2362.62 m a.s.l) and is therefore some 300 m higher in elevation compared to the glacier terminus. The chosen scanning position enabled an excellent overview of the study area with minor shaded areas (Fig. 2). In this study the two different applications of TLS data acquisition were carried out: (i) for the delineation of landforms using up to 6 scanning sectors with very high resolution (DTM resolution 0.20–0.25 m) and (ii) for area-wide calculations using one overall scan (DTM resolution 1.0 m) (Table 1).

GPS – both standard and differential GPS – was used to depict the location of the glacier boundary at the lowest part of the glacier tongue at Pasterze Glacier almost annually since 2003 (Avian et al., 2007; Kellerer-Pirklbauer, 2009). For this study GPS data sets of every epoch between 2010 and 2013 were considered (Table 1) and used to verify and support the delineation of glacier boundaries on the basis of the TLS-DTMs. Although TLS campaigns and GPS measurements were not carried out exactly at the same time, consideration of this in situ evidence further enhanced the decision process in delineating the glacier boundary.

3.2. DTM generation and quality control

After processing in Riegl RiScanPro all point clouds were transferred to Surfer 10. This software package enables user-defined DTM-
generation and effective error management. Generally the quality of DTM
generated by (i) the acquisition method and processing procedure (laser system, processing method and interpolation algorithms) as well as (ii) the surface and hence data characteristics (e.g. reflectivity of surfaces, footprint, GSD). As a consequence two levels of geometric resolution of the raster cell were calculated, namely (i) 0.2 as well as 0.25 m for visual landform interpretation and (ii) 1.0 m for calculation purposes within the DTM of differences (DoD) due to the large extent of the study area (Table 1).

In order to determine the surface representation uncertainty \( \delta(z) \), we used a maximum of twelve distinct reference areas per epoch, where seven reference areas were expected to show no changes (for quality analysis between two epochs) and five reference areas were expected to show significant changes in surface elevation (to improve the quality analysis at one epoch). These areas were located at (i) different distances and (ii) showing different surface characteristics including bedrock and consolidated and stable debris areas (no change) as well sanders and debris covered dead ice parts (no changes in the epoch, changes in the observation period). Table 2 shows the characteristics of the measurements of the reference areas.

One approach to characterize the \( \delta(z) \) is to repeatedly measure areas which are expected not to change (Brasington and Smart, 2003). At every epoch each reference area (Table 2) was measured at least two times and the point clouds were subsequently analyzed as follows: (i) Aggregation of all point clouds of all reference areas at one epoch representing the characteristics of a measurement year (Fig. 3/A) and (ii) aggregation of all point clouds of one single reference areas of all epochs representing the long term characteristics of the latter (Fig. 3/B). As individual errors \( \delta z_{new} \) and \( \delta z_{old} \) of the corresponding DTM propagate into the DoDs, Brasington et al. (2003) provided the relation

\[
\delta U_{DoD} = \sqrt{(\delta z_{new})^2 + (\delta z_{old})^2}
\]

(1)

to determine the propagated error \( \delta U_{DoD} \) in the DoD. In assessing the significance of DoD uncertainties we used the standard deviation of the error of each cell in estimating the \( \delta(z) \). In doing so Eq. (1) is modified to (Wheaton et al., 2010):

\[
U_{cit} \sim t \sqrt{u_{cit} u_{cit}}
\]

(2)
where $U_{\text{crit}}$ depicts the critical threshold error, $t$ is the critical t-value at the chosen confidence level.

$$t = \frac{Z_{\text{DTM}_{\text{new}}} - Z_{\text{DTM}_{\text{old}}}}{\sqrt{\sigma_{\text{z}_{\text{new}}}^2 + \sigma_{\text{z}_{\text{old}}}^2}}$$  \hspace{1cm} (3)

In order to determine quality parameters, all measured points of one epoch were aggregated to a new point cloud. The distribution of original points within the reference areas bedrock1 and debris (composition of the three measurements of each reference areas in 2012) can be seen in Fig. 3/C. Following the approach presented above $U_{\text{crit}}$ of corresponding DoDs was calculated to be between 0.2 cm and 17.5 cm. Therefore all measurement values smaller than 20.0 cm were not considered in the following analysis (Table 2).

### 4. Visual landform classification

Following Ballantyne (2002), a visual landform classification forms the basis for surface interpretation of geomorphological systems and sub-systems in the proglacial zone. All landforms in the study
area were classified visually applying a self-developed hierarchical interpretation key (Table 3). This interpretation key consists of 11 different landforms belonging to three different landsystems, one having two subgroups. The classification process was accomplished manually in ArcMap (2D) and in ArcScene (2.5D) using the DTMs. In order to support the classification we furthermore used orthophotographs (2012) and terrestrial photographs taken during several field verification campaigns.

Differentiating between the glacier foreland system and the glacier system is not trivial as a result of the widespread supraglacial debris cover. Exceptions are ice cliffs in the debris-covered part of the glacier. Thermo-erosional notches develop sometimes at the foot of such ice-exposed slopes if a stagnant water body is adjacent to it (Kellerer-Pirklbauer, 2008). On the basis of our multi-temporal DTMs we were able to differentiate between ice-marginal areas with significant surface lowering and ice-marginal areas with minor surface lowering. The boundary between these two areas was commonly very distinct and was subsequently used as the glacier/non-glacier boundary (cf. Kaufmann et al., 2015).

5. Results

5.1. Temporal evolution of landform types

The total area of the landform classification covered 0.885 km² in the three epochs 2010, 2012, and 2013. The results of the landform classification in the study area show a substantial increase of the glacier foreland landsystem – both in the depositional and the erosional...
In terms of vertical surface elevation changes and glacier delineation, the entire proglacial area was analyzed separately for each year. Hence, the proglacial area is not equal for every period. The analyzed proglacial area increased from 0.127 to 0.214 km² during the period 2010/13. With respect to vertical surface elevation changes the study area can be divided into three substantially different geomorphological environments by DTM comparison and visual landform interpretation (Fig. 5). The first area, the environment of very recent (at least superficially) deglaciated terrain is characterized by subsurface ice melting causing extensive surface subsidence and consists primarily of sandur and terrace landforms. This type is subsequently termed here as the “valley bottom area”. Second, we defined an area including the north-eastern hillside until the distinct break of slope as the “footslope area”. This area is also characterized by surface subsidence (although moderate) and mainly consists of extensive dead ice bodies covered by debris and former subglacial river beds. Third, the environment of the adjacent and higher elevated lateral slope area was delineated to comprise mainly steeper drift-covered slopes (partly with terrace structures) with incised gullies. This area is subsequently termed as the “lateral slope area” (Fig. 6). The aerial increase of the proglacial area which resulted from glacier terminus recession was accompanied by a total volume loss of approx. 1,309,000 m³ (±37,500 m³), or a median vertical surface elevation change of −3.44 m during the entire observation period. This represents a combination of glacier ice melting at the surface and below the surface (buried dead-ice and ice still connected to the main glacier body below the surface) and downvalley sediment transfer. Removing the glacier domain and solely considering the glacier foreland system on an annual basis of every epoch, the volume loss was calculated to be approx. 275,000 m³ (±22,200 m³) for the entire observation period. The median of vertical surface elevation change varies from −0.72 (2010/11) to −0.26 m in the observation period 2012/13 the period 2010–13.

5.3. Surface elevation changes: distinct areas

Our analyses revealed several areas with distinct surface changes which are discussed in more detail further below. Seven such areas of distinct surface elevation change are located in the valley bottom area, six in the footslope area, Fig. 6 depicts the location; Table 5 numerically lists the characteristics of these areas.

5.3.1. Foothslope area

Special area drift mantled slope 1 (Fig. 6, DMS-1): In detail at the footslope area a large area with significant negative vertical surface elevation changes can be detected in all DTMs. This special area DMS-1 (Table 5) shows negative surface elevation changes in every epoch which were significantly higher than in the adjacent slope area. As a zone of aerial erosion and gully formation the special area DMS-1 lost >35,000 m³ in the period 2010/13 and decreased simultaneously in area from 21,000 to 9700 m² in the same period. The median vertical subsidence also decreased from −0.75 m a⁻¹ (2010/11) to −0.48 m a⁻¹ (2012/13) during the observation period.

Special area gullies (Fig. 6, GUL-3): The analysis of surface elevation differences were only carried out in the entire observation period 2010/13 due to significance reasons (see Table 2). The results revealed three gullies with negative volume changes in the entire observation period. The formation of GUL-0 and GUL-1 indicates an active erosion channel, GUL-1 was divided into sub-units for interpretation reasons. All three observed gullies were in a state of ongoing activity expressed by an overall volume loss of −380 m³ (GUL-1a) and −341 m³ (GUL-1c) and a distinct volume increase of 345 m³ in the deposition section (GUL-1b) between 2010 and 2013. Below the special area DMS-1, two gullies (GUL-2, GUL-3) represent the link to the valley bottom area. GUL-2 showed a volume loss of −1400 m³.
5.3.2. Valley bottom area

In the overall observation period 2010/13 a volume of ~3350 m$^3$ deposited in an increasing area (SAN-1) of 850 (2011) to 3600 m$^2$ (2013). A second area of deposition (SAN-2) developed in 2012 with 2900 m$^2$ and an increase of volume of 1400 m$^3$ in the period 2012/13.

6. Dynamics of recent geomorphic processes

The very fast recession of the Pasterze Glacier in the last three decades caused a vast area of sander, dead ice bodies and drift mantled slopes. This study tries to characterize this area addressing different process types by identifying different landforms and interpreting vertical surface elevation changes. Following the approaches of Ballantyne (2002), Mercier (2008), and Slaymaker (2011) the location of processes in the Pasterze Glacier area can be distinguished in proglacial processes and paraglacial processes (Fig. 8). Recent glacial processes, processes related to past glaciation, processes related to ground ice and subsequently to phase changes of water – all resulting in sediment mobilization – have a strong impact on geomorphological activity (Beylich et al., 2006). Carrivick et al. (2013) noted that in the nearby catchment of the Ödenwinkelkees (10 km NW of Pasterze proglacial area) process-chains can be roughly distinguished between (i) the erosion of lateral slopes by gravitational processes and (ii) the construction of landforms by fluvial processes. Contrary to other large alpine glaciers such as Belvedere glacier (e.g. Mortara et al., 2009) or Lower Grindelwald Glacier (Stoffel and Huggel, 2012), Pasterze Glacier has not formed prominent lateral moraines since the LIA. Consequently there are no large ice cored moraines (e.g. at the Athabasca Glacier, Alaska; Mattson et al., 1993) in the study area that would make very large sliding events as well as collapses possible (Chiarle and Mortara, 2001). In the slope parts of the study area of Pasterze Glacier, the dominant processes are linear erosion, debris-flows forming gullies and subsequent sedimentation of the remobilized sediments. The sedimentation occurs further below gully formations in the form of debris-flow cones deposited atop of former lateral morainic sediments or kame-terraces (Fig. 9). Dead-ice is most likely absent in this part of the study area and therefore changes in surface elevation are solely connected to erosion and sedimentation processes. This evidence is in contrast to the lower part of the study area comprising the valley bottom and part of the more gently-slope hillslope area, where dead ice is widespread. Depending on the thickness of the sediments, the magnitude of subsurface ice melting varies substantially between locations. For the tongue of Pasterze Glacier Kellerer-Pirklbauer et al. (2008) revealed that ablation is reduced by 30–35% under a debris cover of at least 15 cm thickness.

In the proglacial area of Pasterze Glacier the thickness of debris-cover is generally several decimeters as indicated by significant field evidence (e.g. Fig. 7/1). In considering (i) a realistic minimum debris thickness of 50 cm in this area (ii) a yearly mean ablation rate of 8.6 m (see above) at nearby debris-free surfaces of the glacier, (iii) an absence of permafrost conditions at the glacier tongue (Kellerer-Pirklbauer et al., 2012), and (iv) the non-linear relationship between debris thickness...
and ablation rates (e.g. Østrem, 1959; Mattson et al., 1993), one might estimate a yearly sub-debris ice melting rate in the order of 1 m a\(^{-1}\). This rate is, however, substantially higher at steep bare-ice slopes (i.e. slope of terraces) where shortwave radiation influences the ice-surface heat budget (Han et al., 2010). Sedimentation rates in the proglacial area are substantially lower compared to dead-ice melting. Therefore, for the entire proglacial part of the study area a net negative mass balance of approx. 275,000 m\(^3\) in the entire observation period is evident (mean 90,000 m\(^3\) a\(^{-1}\)). In the Eastern Alps other studies show similar results such as >80,000 m\(^3\) a\(^{-1}\) at recently deglaciated areas at the Gepatschferner (Baewert and Morche, 2014). Carrivick et al. (2013) revealed 4400 m\(^3\) a\(^{-1}\) at the smaller proglacial area at Ödenwinkelkees.

However in detail, particular landforms of accumulation were detectable especially as several sequences of debris flow cones below gully formations in the “footslope area” and minor depositional areas in the sandur area. Net increase in surface elevation or deposition of sediments at the proglacial area of Pasterze Glacier most likely begins at the “Sandersee” (Fig. 2) at a distance of c. 250 m downvalley of the study area (Geilhausen et al., 2012).

Within the study area the “lateral slope area” is characterized by drift mantled slopes and is partly affected by linear erosion. Valley-side streams do not play a major role for sediment transport within the study area (Ballantyne, 2002); there are no active channels with perennial channel flow at the lateral slopes. At the area of the transition to a lower slope gradient we detected a large number of active gullies and corresponding debris cones. The “footslope area” is dominated by sequences of ice-contact sediments and originally subglacial or ice-marginal channels. The landforms within the valley bottom area in very recently exposed areas are characterized by a chaotic pattern of irregular hummocks and hollows formed by melting debris covered ice bodies which are typical for receding glaciers (Hubbard and Glasser, 2005; Geilhausen et al., 2012). In this area thermo-erosion and ablation extensively mobilized a large amount of sediments affecting the dominant landform group of ice-cored terraces and ridges built of glacio-fluvial and glacial sediments. Generally extensive erosion events mainly occur at recently deglaciated areas around the glacier terminus and are rapidly followed by depositional events. The degree of activity of these processes is reduced downvalley as shown by a downvalley decrease of process magnitudes and process diversity (Geilhausen et al., 2012).

Fig. 5. Spatial distribution of landforms in the study area Pasterze proglacial area in the epochs 2010 to 2013. Redline indicates glacier boundaries. Codes in the legend are described in Table 4. Shaded relief in the background is for orientation purpose.
6.1. Process dynamics in the lateral slope area

The gully structure in the upper part of the “lateral slope area” is located beneath the extensive rock walls building up the Freiwandkopf-ridge. The rock walls contain large channels which act as a linear drainage system transporting the precipitation further down to the foot of the rock walls (at approx. 2350 m a.s.l.), where the LIA moraine is located immediately beneath the tourist track “Gamsgrubenweg” (Fig. 8/2). This situation favors the formation of debris flows within the drift material. The lateral LIA moraine is deeply incised, creating a formation of gullies (Fig. 8/2). Furthermore drift which is accumulated at the gully floors becomes remobilized (Ballantyne, 2002).

Fig. 6. Vertical surface elevation changes during the periods 2010/2011, 2011/2012, 2012/2013, and 2010/13 (Detail “A”). Shaded relief calculated from corresponding annual DTMs. Surface elevations changes of ±0.175 m are not considered due to accuracy restrictions (Table 1). Codes enclosing the special areas of surface elevation changes correspond to Table 5.
Rickenmann and Zimmermann (1993) stated that such unvegetated slopes – consisting of glaciogenic material with an inclination of 27° to 38° – form important source areas for debris flows. Ballantyne and Benn (1994) showed that gullying of upper drift-mantled slopes result in a reduction of the mean slope angle from 35° to 30°. Curry et al. (2006) furthermore found that modified slope profiles converge to an equilibrium form. This is characterized by an upper rectilinear slope of 29° ± 4°, also termed as a minimum gradient of hillslope debris flows (e.g. Takahashi, 1981; Rickenmann and Zimmermann, 1993). The “lateral slope area” (Fig. 6) shows a median slope angle of 35°.8° which exceeds the limit of Curry et al. (2006) of 29° ± 4° and can therefore be termed as an over-steepened drift-mantled slope. The high slope gradient might be explained by a rather thin, well-compacted cliff-forming (Fig. 7/C) layer of drift material, with a maximum thickness of several meters, above steep bedrock.

The upper gully formation inside the LIA area downslope ends where the median slope gradient decreases from 36° to 25° (Fig. 8/F; see also Curry et al., 2006) apart from the areas where bedrock is exposed at the surface (Fig. 7/E). Furthermore, Curry et al. (2006) stated that gullies appear to reach their maximum depth and width ca. 50 years after deglaciation. This observation can be generally confirmed by our observations for the drift gullies at the upper limit of the slope below the LIA limit (Figs. 7/B, 8/F). Solid vegetation cover at the upper part and stabilized gully sides give clear evidence of no substantial further lateral extension of the gullies (Fig. 7/B) in recent times. However, the gullies themselves show unconsolidated debris and are obviously active transit tracks from the bedrock above the LIA lateral moraine to the lateral slope area further below.

6.2. Process dynamics in the footslope area

Generally major slope adjustment processes within the “footslope area” occur in the slope transition zone (Fig. 8/F) until the lower margin at the ice-cored terraces (Fig. 8/D). In terms of surface morphology the “footslope area” is mainly characterized by sequences of former ice-contact sediments building terrace-like flattennings in the slope (Figs. 7/C, 8/F, Fig. 9, stage 1). These formations were developed in stages of fast lateral recession following stages of slower lateral glacier recession where sediments were deposited. These ice-contact terraces within the drift-mantled slopes can be related to glacier stages (Fig. 7/D, E, F) over the entire slope. At the transition zone of the changing slope-gradient (Fig. 8/F) deep gullies incised into these ice-contact terraces. The activity of these gullies are expressed by an increase of the area of the gully formation (Fig. 6, GUL 1-2) and can both be explained (i) by an upslope elongation of the gullies themselves and (ii) an corresponding enlargement of the gully profiles in active sections (Fig. 7/B). At the foot of the active gullies debris cones were formed as a consequence of accumulation processes (Fig. 8/F). However, the ice-contact terraces with the paraglacial debris-flow sediments atop are frequently linearly eroded by incision through backward wasting (Fig. 9, stage 2). Younger events buried older flow deposits and produced a highly disturbed micro-topography (Fig. 9, stage 3). Whether these flows are connected to extreme events or frequent events has not yet been clarified. This reworking of lateral morainic material forms an exemplarily process in recently deglaciated areas and was reported by several authors (e.g. at the Glacier du Mont Miné; Curry et al., 2006).

In the most northeastern section of the “footslope area”, a large area (Fig.,6, DMS-1; Fig. 8/F) with significantly higher negative vertical surface elevation changes was detected during the observation period. This area is expressed by a noticeably different surface texture where a brighter sediment color indicates active slope processes. Furthermore, “dirty” seasonal snow patches which are remnants of debris-charged snow avalanches are frequently found until the month of June at this site (elevation 2110 m a.s.l., aspect SW). DMS-1 can be characterized as (i) underlain by remnant dead ice bodies and (ii) under the influence of very active slope processes. These processes occur due to (i) the vicinity of the glacier (deglaciation probably 5 years ago) and slope characteristics, and (ii) the location of the ice-margin and (iii) dead ice evidence detected in field work.

In the lower part of the “footslope area” ice marginal channels play a significant role in sediment transport. These channels show large floors and cliff-forming limits of >10 m (Fig. 7/E and F). Sediments consisting of glaciofluvial (and glacial) deposits at the floor of these former meltwater channels are mostly rounded and partly separated from the recent drainage pattern (Geilhausen et al., 2012).

6.3. Process dynamics in the valley bottom area

The entire valley bottom area underwent a massive change in surface characteristics during the observation period. A total volume of approx. 275,000 m³ (±22,200 m³) was lost in the valley bottom area between 2010 and 2013. This cannot be explained only by downwasting of buried ice, but also includes mobilization and fluvial reworking of sediments leading to extensive proglacial sediment transport and landscape modification (Ballantyne, 2002). However, in the interpretation of this comparably large volume value we have to consider the influence of different water levels during the measurements. TLS campaigns were carried out mainly in the forefront to avoid the effect of the diurnal variation of the water level. In comparing images acquired during measurements, a significant difference in the area covered by water and the water level itself could be detected between the single epochs. Although TLS generally does not get reflected impulses from water, the existing turbidity of glacial water attenuates this effect and an acceptable amount of pulses can be acquired.

Generally it can be stated that the water areas at the valley bottom area were significantly increasing in terms of spatial extent during the observation period. In 2010 large areas were covered by water, with a minor extent of dammed water areas from glaciofluvial deposits, and in 2012 large pond-like kettle water areas (Fig. 8/I) were visible. The channel pattern within the sandur/water area is highly variable over the observation period (Morche et al., 2008). In ice-proximal locations dead-ice holes (DIH) are very frequent landforms. In detail we detected the co-existence of fast re-filling dead-ice holes at flat debris covered areas and ongoing extensions of already large dead-ice
holes at the sandur area close to the glacier margin (Fig. 8/13). Ice-cored landforms (TER, TES, Fig. 7/D) are the most dominant and dynamic landform types in the valley bottom area. The surfaces of these landforms frequently consist of a fluvially deposited layer of subrounded to rounded gravel with sand (Geilhausen et al., 2012). Thermo-erosion of these ice-cored landforms also results in significant sediment reworking (Ballantyne, 2002; Kellerer-Pirklbauer, 2008). Deposits mostly consist of coarse-grained diamictons with sandy to gravelly matrix (Geilhausen et al., 2012) and layered sediments depending on the acting process. The evidence of re-filling of recently formed hollows and mobilization of very young sediments agrees with Carrivick et al.’s (2013) findings in the nearby proglacial area of Ödenwinkelkees Glacier. In the valley bottom area, especially the extensive areas of DIB (Fig. 6) at the northeastern margin of the glacier terminus as well as the terrace landforms, (TER, TES) places of mass-wasting processes (Etzelmüller, 2000) were found. Such events occurred at locations where the dead ice was exposed. Especially at the TER sites (TER-1, TER-2, and TER-3 in Fig. 6), slope failures exposed buried ice and accelerated the melting process. Several sediment slide events per day were observable during measurements mostly occurring during the afternoon. Both DIB landforms (DIB-1, DIB-2 in Fig. 6) lost a volume of approx. 40,000 m³ in one year mostly through the melting of sub-sediment glacier ice. Due to the close vicinity of two well-frequented tourist tracks these events at dead-ice landforms also constitute a serious natural hazard.

On the other hand, particular terraces and slopes of ice-cored terraces (TES) show massive retreat in meltwater proximal locations and contribute to downvalley sediment transfer. The slope gradient
between the glacier terminus area and the sandur lake is very low, causing a reduced transport capacity for coarser sediments. Therefore this area is very well suited for the deposition of clasts of any size. In the lower part of the valley bottom area, zones of sedimentation were detectable which could be considerably higher than the measurement values due to constant subsurface ice melting.

Fig. 8. Top: Central part of the study area "Pasterze Glacier proglacial area" with the rock faces of Freiwand and Franz-Josefs-Höhe at the debris covered glacier part. Bottom: Schematic profile A-B. Numbers: (1) uppermost gullies in the rock walls of the mountain Freiwand, (2) tourist track "Gamsgrubenweg", mostly identical with the root zones of the gully structure in the image, (3) LIA maximum partly expressed by a lateral moraine, (4) lateral gully edges, (5) transition zone from steep slope to moderate slope = boundary between "lateral slope area" and "footslope area", (6) partly ice-cored ice-contact sediments with date of formation, (7) foot of slope, (8) terminus area of the "clean ice" glacier part, (9) boundary of the "debris covered" glacier part, (10) sediment part of sandur area, (11) water part of sandur area, (12) approximate location of upslope limit of significant vertical surface elevation changes = boundary between "footslope area" and "valley bottom area" (13) dead ice holes. A and B form the profile shown below. TLS position is indicated at the Franz-Josefs-Höhe at the top-right corner.
Extensive areas prone to paraglacial processes are the result of rapid deglaciation in climate sensitive alpine catchments. The evidence within the study area at Pasterze Glacier shows a typical example of landform evolution and modification due to a fast retreating valley glacier exhibiting different magnitudes of processes and levels of maturity of landforms. This study provides results from high resolution TLS-data including (i) landform classification and (ii) vertical surface elevation changes on an annual basis between 2010 and 2013. In doing so we did not only analyze single landforms but also focused on a spatial fragmentation of geomorphological activity (Carrivick et al., 2013) in differentiating the study area into varying “areas”.

All observations, such as landform classification and patterns of vertical surface elevation changes, led to the differentiation of the entire study area into (i) a lateral slope area, (ii) a footslope area, and (iii) a valley bottom area based on dominating geomorphic processes. Furthermore results do not only confirm that recently deglaciated areas are subject to high geomorphic activity, but that these landforms might change their shape and sizes substantially from one year to the next. Especially processes at ice-marginal locations are not only dominated by erosional/depositional processes, but also the melting of buried dead-ice bodies plays a major part within the process chain.

In the study area the lateral slope area is mainly characterized by consolidated drift mantled slopes incised by linear erosion forming large gully structures. The significant change in slope gradient forms the transition to the footslope area and is accompanied by an increased occurrence of subsurface ice melting which is still a dominant process. Erosional processes such as extensive incision occurring at former ice-contact sediments (representing different glacier stages) are also widespread. The upper part of the footslope area as well as the lateral slope area are characterized by a rather thin layer (up to several meters) of debris covering steep bedrock. In this domain linear erosional processes seem to dominate which transport a significant amount of sediments to the footslope area. Furthermore, substantial deposition of sediments from the upslope gullies occurred at the surface of slightly inclined ice-contact sediments forming debris cones. At the lowest section of the study area nearly the entire valley bottom area is expressed by extensive surface subsidence related to dead-ice melting and – to a minor extent – sediment transport out of the study area into the sandur lake and beyond.

Some of our observations suggest an enhancement of the temporal resolution of the acquisition of TLS data concerning monthly measurements beginning in June to September (i.e. Baewert and Morche, 2014). This effort would lead to a very useful data basis for analyzing changes at (i) the water and sandur area due to different discharges within a season (Milan et al., 2007), at (ii) the footslope area at different process magnitudes due to the varying availability of surface and subsurface (melt-) water. Furthermore, to get a better understanding of subsurface ice conditions and volumes, repeated geophysical measurements (e.g. electrical resistivity tomography) will be considered in the future (Bosson et al., 2014). In terms of methodology TLS measurement strategy has already been enhanced substantially to optimize results, as not all of our high-resolution TLS data cover the entire study area. As a consequence we established two new scanning positions in the study area in September 2013. Consequently we want to take a closer look at the development of the debris mantled slope by monitoring e.g. the development of the upslope gullies (e.g. considering analysis about gully density as in Curry et al., 2006) and the evolution of the downslope sediment areas.

Acknowledgements

We are very thankful for the support of several students of the University of Graz as well as the Graz University of Technology. Furthermore we want to thank Daniel Wujanz and Matthias Rathofer for their support during measurements and post-processing (Graz University of Technology). Prework to this study was carried out within the framework of the projects ‘ALPCHANGE – Climate Change and Impacts in Southern Austrian Alpine Regions’ financed by the Austrian Science Fund (FWF) through project no. FWF P18304-N10 as well as ‘PermaNET – Permafrost long-term monitoring network’. PermaNET was part of the European Territorial Cooperation and co-funded by the European Regional Development Fund (ERDF) in the scope of the Alpine Space
Programme (project no. 18–1–3–I). We would also like to recognize the company GOHAG for providing valuable logistical support. Finally we highly appreciate the valuable comments and suggestions of the anonymous reviewer.

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