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Contemporary geomorphological activity throughout the proglacial area of an alpine catchment

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ABSTRACT

Quantification of contemporary geomorphological activity is a fundamental prerequisite for predicting the effects of future earth surface process and landscape development changes. However, there is a lack of high-resolution spatial and temporal data on geomorphological activity within alpine catchments, which are especially sensitive to climate change, human impacts and which are amongst the most dynamic landscapes on Earth. This study used data from repeated laser scanning to identify and quantify the distribution of contemporary sediment sources and the intensity of geomorphological activity within the lower part of a glaciated alpine catchment; Ödenwinkelkees, central Austria. Spatially, geomorphological activity was discriminated by substrate class. Activity decreased in both areal extent and intensity with distance from the glacier, becoming progressively more restricted to the fluvially-dominated valley floor. Temporally, geomorphological activity was identified on annual, seasonal, weekly and daily timescales. Activity became more extensive with increasing study duration but more intense over shorter timescales, thereby demonstrating the importance of temporary storage of sediment within the catchment. The mean volume of material moved within the proglacial zone was 4400 $m^3.yr^{-1}$, which suggests a net surface lowering of 34 $mm.yr^{-1}$ in this part of the catchment. We extrapolate a minimum of 4.8 $mm.yr^{-1}$ net surface lowering across the whole catchment. These surface lowering values are approximately twice those calculated elsewhere from contemporary measurements of suspended sediment flux, and of rates calculated from the geological record, perhaps because we measure total geomorphological activity within the catchment rather than overall efflux of material. Repeated geomorphological surveying therefore appears to mitigate the problems of hydrological studies underestimating sediment fluxes on decadal-annual time-scales. Further development of the approach outlined in this study will enable the quantification of geomorphological activity, alpine terrain stability and persistence of landforms.

KEYWORDS sediment flux; landscape denudation; LIDAR-Laser scanning; glacier; river

HIGHLIGHTS

- Quantified distribution and intensity of contemporary geomorphological activity
- Categorisation of geomorphological activity by substrate class
- Inter-annual, seasonal, weekly and daily analysis of magnitude - frequency regime
- Mean volume of material moved within the 9.2 km^2 catchment of 4400 m^3 per year
- Net surface lowering across the whole catchment of at least 4.8 $mm.yr^{-1}$

INTRODUCTION AND RATIONALE

Understanding contemporary sediment fluxes is fundamental to predicting the likely effects of future changes to geomorphological activity and landscape development, whether those changes are induced by climate change or by human activity (c.f. Jones, 2000; Slaymaker, 2010). Catchment-wide denudation is commonly inferred indirectly from rates of fluvial suspended sediment exiting catchments (c.f. Milliman and Syvitski, 1992). However, it is important to recognize that the discharge of suspended sediment *from* catchments effectively considers a catchment as a ‘black box’; it does not represent all of the geomorphological activity that occurs *within* that catchment (Caine, 2004), nor does it recognise the spatial and temporal variability of that activity. This problem has been acknowledged for several decades by projects that have examined bedload movements and that have defined sediment production, transfer and storage within a catchment (e.g. Rapp, 1960; Warburton, 1990; Trimble, 1995).

Future changes to geomorphological activity and landscape development will be especially rapid and potentially severe within alpine catchments because they are very sensitive to climate changes and to human impacts. This sensitivity is most evident in water availability (c.f. Barnett et al., 2005), water quality and stream biodiversity (e.g. Brown et al., 2003; 2007), water thermal dynamics (e.g. Carrivick et al., 2012) and sediment fluxes (c.f. Milliman and Syvitski, 1992; Hallet et al., 1996). Understanding contemporary sediment fluxes *from* alpine catchments has to account for the considerable variability in geomorphological activity between adjacent mountain catchments (e.g. Gurnell et al., 1988; Trimble, 1995; Carrivick and Rushmer, 2009). However, understanding contemporary sediment fluxes *within* alpine mountain catchments is complicated because mountain glacier responses to regional and local climate are heterogeneous in space and time (e.g. Carrivick and Chase, 2011) and because there is often a significant imbalance between sediment production and sediment transport due to former glacial activity that i) over-steepens topography and promotes paraglacial slope adjustment processes, ii) produces large sediment stores available for erosion, and iii) emplaces moraines that can be both a sediment source and a barrier to meltwater (Beylich and Warburton, 2007). This complexity in the spatial and temporal nature of geomorphological processes hinders the identification and quantification of sediment sources, storages and fluxes (e.g. Dietrich and Dunne, 1978; Jones, 2000; Bertoldi, et al. 2009). For example, it is well known from hydrological measurements of suspended sediment that small mountain catchments have a particularly variable sediment flux that is seldom resolved, partly because large but short-lived events are often missed (Kirchner et al., 2001; Lewis et al., 2005). Consequently, decade-long sediment-yield measurements using conventional (hydrological) methods can greatly underestimate long-term (centennial – millennial) average rates of sediment delivery (Kirchner et al., 2001). Short-term geomorphological activity within parts of a catchment can be determined from repeated topographic measurements and episodic sediment fluxes can be calculated as a volume of material moved between each of these surveys (e.g. Martin and Church, 1995; Ham and Church, 2000; Fuller et al., 2003). However, whilst several European alpine countries are in the process of making systematic country-wide Airborne Laser Scan surveys, use of ALS and Terrestrial Laser Scan (TLS) topographic data (i.e. Light Detection and Ranging; LiDAR data) to determine geomorphological changes within alpine catchments is presently limited. This is perhaps because ALS datasets tend to be acquired on a campaign basis, rather than as part of routine monitoring strategies. It is also undoubtedly because of the problems of processing such voluminous and complex datasets.

The overall aim of this paper is to identify and quantify the contemporary distribution and intensity of activity of sediment sources, storages and fluxes within the proglacial part of a glaciated alpine catchment.

QUANTIFYING GEOMORPHOLOGICAL CHANGES WITHIN ALPINE CATCHMENTS

Long-term (centennial-millennial) sediment storage within alpine catchments has been quantified by combining geophysical surveys, digital topographic analyses and geographic information system (GIS) modelling techniques (e.g. Otto et al., 2009; Schrott et al., 2003). Determination of contemporary sediment sources, storages and fluxes within alpine catchments remains problematic however, not least because existing catchment-wide models (e.g. Caine, 1974; Dietrich and Dunne, 1978) are qualitative. These qualitative conceptual models are relied on heavily for designing contemporary field sampling of water and sediment fluxes. This is a major drawback with sediment budget studies because rigorous definition of sediment storages and fluxes is necessary prior to a field campaign (Warburton, 1990). Furthermore, it is difficult to decide how to focus field campaigns because sediment storages and fluxes vary greatly over the short-term (annual-decadal) (e.g. Trimble, 1995) due to; i) functional activity of geomorphological coupling is dependent on sediment availability and triggering events (Schrott et al., 2006), and; ii) because intermittent valley-floor and braidplain storage is very important (e.g. Warburton, 1990; Orwin and Smart, 2004; Bertoldi, et al. 2009).

The best way to quantify contemporary geomorphological activity within alpine catchments; and specifically to discriminate contemporary sediment storages and fluxes in space and time, is to employ a geomorphological approach (i.e. to re-survey topography; e.g. Martin and Church, 1995; Ham and Church, 2000; Fuller et al., 2003; Bertoldi et al., 2009). Indeed Orwin et al. (2010) recommend resurveying as the most appropriate method for establishing integrated sediment flux studies in cold environments on inter- and intra-annual time-scales. Re-surveying using traditional methods is exceptionally time-consuming and financially expensive for anything more than a few fixed cross-sections of valley profiles. Differential Global Positioning Systems (dGPS) have helped to alleviate these problems slightly and Schrott et al. (2006) made excellent use of photogrammetric methods to determine changes in sediment storages over a four year period within a deglaciated valley in Germany. Advancements in surveying technology of LiDAR; primarily in the form of ALS and TLS, for rapid very high-resolution analyses (Abermann et al., 2010) have yet to be exploited for holistically examining multi-scale sediment fluxes within highly dynamic alpine catchments.

LASER SCANNING OF ALPINE GEOMORPHOLOGY

High resolution (~ 1 m) topographic data from photogrammetry (e.g. Schrott et al., 2006) and satellite image datasets from mountainous and alpine catchments have to date been used for i) geomorphological mapping, ii) landform unit-scale analyses of episodic geomorphological changes, and iii) analyses of river reach-scale changes (Wang et al., 2010; Smith and Pain, 2009). ALS and TLS instruments give high resolution (< 1 m), high precision (> 0.2 m), and rapid acquisition of surface elevation data over a range of spatial scales and are thereby revitalising geomorphological studies.

Alpine catchment-wide use of ALS and TLS datasets is still new and developing, but a notable work to date is that of Van Asselen and Seijmonsbergen (2006) which illustrated that 1 m resolution Digital Elevation Models (DEMs) can be analysed to map mountain hillslope and elevation properties semi-automatically using object-oriented segmentation and classification techniques. Glaciers have received special attention for monitoring and measurement of retreat, downwasting, and surface character due to the obvious rapid responses to, and consequences of, climate change (Abermann et al., 2010). At a geomorphological unit (i.e. ‘landform’) scale (over tens of metres) and in terms of episodic event-based analyses, Morche et al. (2008) used TLS data to quantify and explain changes on an alpine talus cone within an alpine catchment over a four month period. Dunning et al. (2010) and Abellan et al. (2010) have investigated landslide occurrence and properties. At a river reach scale, repeat surveys using photogrammetric (e.g. Luchi et al., 2007), differential Global Positioning System (dGPS) (Brassington et al., 2000, 2003) and remote sensing data (e.g. Lane et al., 2003) have been used to quantify changes. Hetherington et al. (2005) and Milan et al. (2007) used the same data from a 10 day period in early ablation season (June) to quantify a major episode of avulsion and medial bar erosion as well as transient bank accretion. However, to date no studies have made repeated and multi-scale laser scan surveys within an alpine catchment to identify and quantify the distribution and intensity of contemporary (multi-scale) geomorphological activity and thus sediment sources, storages and fluxes, holistically.

STUDY AREA DESCRIPTION

The Ödenwinkelkees catchment extends from ~ 47^{o}6’00” – 47^{o}8’7”N and from 12^{o}37’19.5” – 12^{o}40’20”E and is partially within the Hohe Tauern National Park, central Austria (Fig. 1). It is well known both for its proximity to the Rudolfshutte Alpinzentrum and for the long-term measurements of the snout position of the Ödenwinkelkees and of the nearby Sonnblickkees (e.g. Slupetzky, 1997; Slupetzky and Aschenbrenner, 1998). The Ödenwinkelkees catchment has an area of 9.2 km^2 and the glacier presently occupies 1.8 km^2 or 19.5 % of that area (Fig. 1A). Catchment terrain surface elevations range from 1790 – 3490 m.a.s.l. (Fig. 1B). The Ödenwinkelkees catchment is composed predominantly of granitic gneiss bedrock (Höck and Pestal 1994) but the hillslopes and valley floor have a superficial veneer of late Holocene (Little Ice Age) and modern scree, moraine and colluvial, alluvial and fluvial sediments (Slupetzky and Teufl, 1991), which are mapped in Figure 2A. The Holocene (de)glacial history of the Ödenwinkelkees catchment is delimited by dated moraines, which are also located in Figure 2A. Figure 3 is an oblique photograph viewing south-south-eastwards from a position very close to the ‘Hinterer Schafbichl’ (Fig. 1A) and it illustrates the catchment ‘mountain landsystem’; specifically the geomorphological coupling between rock faces, hillslopes and moraine ridges and the valley floor. The (de)glacial history (Fig. 2A) and this spatial distribution of geomorphological processes (Fig. 3) led us to target three zones of interest for reporting in this paper; the ‘lower braidplain’, the ‘upper braidplain’ and the ‘proglacial area’ (Fig. 2B).

DATA ACQUISITION AND PROCESSING METHODS

In this study we used Airborne Laser Scanner (ALS) data acquired from July 29th 2008 and multiple sets of Terrestrial Laser Scanner (TLS) data acquired from July and August 2008, August 2009, and June and August 2010 (Table 1). The weather conditions and river discharge during this study period are depicted in Figure 4 and the location of these meteorological and hydrological instruments is given in Figure 1. As an overview, air temperatures were typically below zero on the Eisboden between September and March, rainfall was approximately evenly distributed all summer, and river discharge had a baseflow of ~ 2, a mean of 4, and peaks of 6 – 8 m^3s^{-1} (Figure 4). More details of these meteorological and hydrological data are given in Dickson et al. (2010).

AIRBORNE LASER SCANNING

The ALS used in this study was a Riegl Q560, which is a medium-footprint (~ 0.15 m) LiDAR system. It was mounted onto the underside of a Touring Motor Glider Dimona HK36 TTC-ECO, which operated at ~ 1000 m above the study area terrain. Post-processing of the ALS point cloud data was performed within standard Riegl software RiAnalyse 560. Post-processing incorporated the Dimona onboard dGPS data and its onboard Inertial Measurement Unit data, with dGPS base station data from a Leica GPS500 dual phase receiver located at the Hinterer Schafbichl geodetic control point (Fig. 1A), which is situated at 47°08 04.26241 N, 12°37’41.76277 E. This enabled georeferencing to compute the 3D locations of each of the ALS laser returns within the point cloud. These georeferenced points were then filtered using a combination of text file editing functions to remove atmospheric clouds located altitudinally above the terrain, and then standard zonal ArcGIS functions to remove atmospheric clouds within valleys. The final ALS point cloud had ~ 2 returns per square metre, with elevation values for each point accurate to ~ ± 0.05 m.

TERRESTRIAL LASER SCANNING

We used a Riegl LMS-620 and the Riegl software RiScanPro for both TLS data acquisition and data processing. The maximum field of view (FOV) of this scanner is 360° horizontally and 80° vertically when mounted on a tripod and levelled coarsely to 1 to 2° prior to measurement. Precise levelling was performed during data processing using inclination sensor information with an accuracy of ± 0.008°. To reduce the shadowing effects of large objects and a better visibility of hollows, the scanner was mounted as high as possible above the surface of primary interest; i.e. usually the valley floor. At each scan position (Fig. 2B), a 360° panorama scan with angular resolutions of 0.2 - 0.1° and several high resolution scans with 0.03 - 0.04° resolution were measured leading to 41 single scans. The scanner configuration was specified with 0.15 mrad beam divergence (0.045 cm footprint at 300 object distance) and 'last pulse target' detection mode; which is useful to discriminate between ground surface elements and vegetation features. TLS scan positions and reflector targets, as shown in Figure 2B, were captured four to five times using the same Leica GPS system as used for ALS data acquisition, resulting in an overall precision for TLS data of 2.4 ± 0.7 cm. We were unable to use reflector targets for scans completed from tripod positions at higher elevation on the western valley slopes, so the scanner orientation (towards North) was simply obtained with a magnetic compass (accurate to 1°). This orientation was to provide a coarse georeferencing of each 3D point cloud in the early stage of TLS data processing. Subsequent TLS data processing included accurate georeferencing ('registration') and the elimination of vegetation.

Registration

All scan positions were levelled precisely with the inclination sensor information and georeferenced using dGPS data and tie-point based (reflector) registration. The final TLS and TLS to ALS registrations were achieved by means of the 'Multi Station Adjustment (MSA)', which is a semi-automatic least-square surface matching procedure. In general, surface matching procedures are well established and have been used in different environments delivering good results (Gruen and Akca, 2005; Akca, 2007; Miller et al., 2008). However, TLS to ALS data registration is still a challenge and Bremer and Sass (2012) recently showed that height differences of up to 0.3 m in areas without surface change require further alignments. We therefore integrated the ALS data set in the MSA surface matching procedure. The MSA algorithm divides the raw scan data into square tiles and reduces the number of points per tile by representing them by planes based on defined criterions, e.g. maximum edge length or plane error. If the deviation of the points is too high, the tile is considered not to be a plane and subdivided into sub-tiles until the deviation of the points is within the criterions to define a valid plane. The criteria used in this study were 5.0 m maximum edge length and 0.2 m plane error deduced from several test runs to receive a high number of valid planes representing the scanned surfaces. The spatial orientation and location of all scan positions was then refined in several iterations to achieve the best overall fit by minimizing the normal distance between the planes of overlapping scans from several scan positions and survey campaigns. Thus, all reflectors were included as reference targets, all plane surfaces used were manually revised and areas of significant surface changes between the surveys excluded, effects of surface discrepancies were mitigated through outlier handling and calculation extents were enlarged to provide different spatial orientations of the overlaps. A detailed description of the MSA procedure is given in Riegl (2010). Final MSA results including the standard deviation of distances between all planes used (in cm) as a measure of the final TLS and TLS to ALS registration results are listed in Table 1.

Elimination of vegetation

The removal of vegetation is a crucial processing step as the geomorphological analyses of multi-temporal laser scan datasets should focus primarily on terrain changes and processes. Discrete, sparse and clearly distinguishable vegetation was largely discarded during data acquisition due to the last pulse detection mode. Nevertheless, sophisticated cleaning using the surface comparison functionality within RiScanPro was conducted (Riegl, 2010). Therefore, point clouds were triangulated at different resolutions and outliers representing vegetation were identified by comparing relatively high and low resolution surfaces. However, this could not remove very dense or low vegetation, e.g. alpine meadows and leafy bushes, as identified from our field notes and field photographs. In the case of small-sized features points were therefore deleted manually, but in larger areas this problem remains unsolved. However, the purpose of this study was to quantify geomorphological activity and it was assumed that these areas were inactive due to the presence of vegetation.

DIGITAL ELEVATION MODEL (DEM) CREATION AND ANALYSIS

After processing, the final TLS point clouds were transferred into GIS software using the LAS file format. The quality of TLS derived elevation models is influenced by i) errors caused by the laser system and the applied methodology and algorithms in processing, and ii) the data and surface characteristics, namely point density and type and flatness of the terrain. For this reason, rasterised grids representing the number of points per grid cell were calculated for three point cloud resolutions; 0.2 m, 0.5 m and 1 m resolution and the overall coverage was calculated (Table 2).

Using a subjective consideration of the best resolution-coverage combination (Table 2), both ALS and TLS point cloud data were gridded at 0.5 m cell size resolution to produce a Digital Elevation Model (DEM) for each zone of interest; the ‘lower braidplain’ (0.07 km^2), ‘upper braidplain’ (0.15 km^2) and the ‘proglacial’ zones (1.05 km^2) (Fig. 2B). DEM grid cells without a point within 1 m were returned with a ‘no data’ value. Each zone DEM was clipped to the extent of each substrate class that is listed and mapped in Figure 2A, and as adapted from Slupetzky and Teufl (1991). For convenience and brevity we use shortened versions of the substrate class names throughout the rest of this paper (e.g. pebble-dominated fluvial deposits; river, cobble-dominated alluvial and colluvial deposits; alluvial-colluvial) and we combined the two boulder-dominated substrate classes together; 'boulders'. We differenced DEMs for the same substrate in the same zone for successive surveys and for surveys one, two and three days apart, for the same substrate in the same zone at monthly intervals and the start and end of each summer, and for the same substrate in the same zone at the start and end of the complete study period.

The attribute table of each DEM of difference was analysed to extract the number of grid cells with a given surface elevation change. To distinguish between real and artificial surface changes caused by slightly varying terrain representation due to different point densities and angles of incidences, we calculated DEMs of difference in defined test areas of unchanged surfaces covering approx. 100 - 200 m^2 with different surface characteristics and grain sizes. From these calculations we considered the number of grid cells with surface elevation changes of > 0.15 m (Ø 0.072 ± 0.063 m) to be significant for river, alluvial/colluvial and glacier terrain, and > 0.3 m (Ø 0.17 ± 0.09 m) on boulders and bedrock due to the inherent roughness of these latter surfaces and the greater range of them from scan positions (Fig. 2B). Thus the number of grid cells included in the DEMs of difference gives the number of significant geomorphological events, the area over which these events are occurring and the total volume of those events. The time period between the two surveys defining the DEM of difference permitted the mean rate of volume change to be calculated.

SPATIAL DISTRIBUTION OF GEOMORPHOLOGICAL ACTIVITY

Our descriptions and interpretations of the elevation changes measured by the DEMs of difference are informed necessarily by our knowledge of the study site. In quantification, we interpret negative elevation changes as a result of erosional processes and positive elevation changes as a result of depositional processes. We considered an elevation change in a grid cell to be a geomorphological event and we made no attempt to link adjacent grid cells that had similar elevation changes within the same time period (i.e. to identify distinct landforms).

Geomorphological activity in the lower braidplain zone (Fig. 2B) was predominantly on terrain produced by fluvial processes and alluvial/colluvial processes (Table 3). Fluvial processes produced erosion due to gravel bank collapse (Fig. 5B), erosion due to overbank flooding and winnowing (Fig. 5C) and erosion due to gravel bar migration and cohesive bank erosion (Fig. 5D). Volumes of sediment of up to 23 $m^3 day^{-1}$ (Table 3) were eroded in the lower braidplain zone by these relatively continuous processes. Fluvial processes produced deposition due to overbank sedimentation (Figs. 5B, 5C, 5D) and gravel bar migration (Figs. 5D, 5E), which together comprised ~ -70 m^3 of sediment over three days (Table 3). Irrespective of timescale of observation, it can be seen clearly from Table 3 that geomorphological activity occurred across virtually the whole valley floor in the lower braidplain zone; activity is not discrete. Whilst geomorphological activity that occurred in alluvial/colluvial zones was widespread, the magnitude of activity was highly spatially heterogeneous; episodic erosion and deposition within a gully (Figs. 5E, 5F, 5G) proceeded with volumes up to 97 m^3 per day (Table 3).

The upper braidplain zone (Fig. 6A) was more geomorphologically active than the lower braidplain both in coverage of activity and in magnitude (Table 3). This activity was dominated both in areal extent (up to 94%) and in rate of material (up to 68 m^3day^{-1}) by the active river (Table 3). Overall, this geomorphological activity produced discrete erosion along sections of the braided river gravel banks (Fig. 6B), erosion of a moraine ridge (Fig. 6B), erosion of hillslope (alluvial/colluvial) sediments (Fig. 6C) and widespread erosion of gravel bar surfaces by overbank flooding (Fig. 6D). Unlike the lower braidplain, the upper braidplain appeared to be a zone of net surface lowering; i.e. erosion, with typical mean rates of erosion of ~ 0.5 m^3day^{-1} (Table 3). Deposition across the upper braidplain was predominantly a result of fan apex aggradation (Figs. 6B, 6D) but some discrete positive elevation changes appear to be due to gravel bar migration (Figs. 6B, 6C, 6D). Some of the speckled positive elevation changes in Figures 6B and 6C were undoubtedly due to vegetation growth and were not considered further. The relatively high volumes and rates for the upper braidplain boulder class (one day) (Table 3) were due to melt of snowpatches and so are ignored herein.

Due to its highly variable nature, we draw attention to the fan apex at the head of the upper braidplain part of the river. Within two separate three-day periods this fan experienced either intense channel migration and avulsions, a complete re-organisation of the surface drainage pattern (Fig. 6E), or relative inactivity (Fig. 6F). Within three separate one-day periods the river eroded moraine banks (Fig. 6G), produced widespread fan head aggradation (Fig. 6H) or widespread fan head lowering (Fig. 6I).

Geomorphological activity in the proglacial zone (Fig. 7A) was most widespread and most intensive when compared to the other two areas of the catchment (Table 3). The glacier both retreated slightly and lowered in surface elevation by up to ~ -500 m^3day^{-1} (Table 3) and there were numerous minor surface elevation changes on terrain categorised as alluvial/colluvial (Table 3). Figure 7 depicts a particularly active part of the proglacial river that was both constructing new gravel bars (Figs. 7B, 7D) and eroding gravel bars and gravel river banks (Fig. 7C). The proglacial zone of the river also had net mass loss, typically of -6 m^3day^{-1} and of up to -23 m^3day^{-1} (Table 3). The proglacial river is clearly inactive at some parts of the year; the active area in summer months (50%) and weeks (63%) is less than between years (up to 91%). Terrain classified as bedrock within the proglacial zone had large areal extents of activity, volumes of up to 2600 m^3 and rates of sediment movement of ~ 7 m^3day^{-1} (Table 3) due to episodic geomorphological activity within gullies on the flank of a prominent Little Ice Age moraine ridge (Figs. 7F - 7I), and due to melt of snow patches.

Overall, 60% of the area of the Ödenwinkelkees catchment comprised landforms and land surfaces that are apparently decoupled from contemporary geomorphological systems. Contemporary geomorphological activity was found to be limited spatially to the Eisboden valley floor, which represents 4.6 % of the catchment area, and to adjacent moraine and scree slopes covering 10.9 % of the catchment area. Geomorphological activity within our study period and study area was dominated by 'continuous' low-magnitude processes.

TEMPORAL INTENSITY OF GEOMORPHOLOGICAL ACTIVITY

Whilst we were unable to capture every geomorphological event in the Ödenwinkelkees catchment, we were able to measure the aggregate effects of geomorphological activity between surveys (Table 3). Thus, we can deduce the relative levels of geomorphological activity through the winter months versus the summer months, for example, as we can for activity occurring within monthly, weekly and daily time-scales (Table 3). Larger elevation changes are interpreted to be a result of more intense geomorphological activity.

Across terrain classified as active river, the proglacial, upper braidplain and lower braidplain zones experienced a mean volume change (per grid cell) over the two year study period of -5.1 m^3, -0.5 m^3 and +0.6 m^3, respectively (Fig. 8). Across the winter months of 2009 – 2008 and 2009 – 2010 the proglacial zone experienced a mean volume change (per grid cell) of -3.3 $m^3$and -0.3 m^3, respectively (Fig. 8). In contrast, the proglacial zone had a mean volume change (per grid cell) of 1.3 m^3 in two summer months and in the same two summer months the lower braidplain had 0.04 m^3 (Fig. 8). Over (multiple) three day periods the active river in the proglacial, upper braidplain and lower braidplain zones had mean elevation changes (per grid cell) of 2.4 m^3, -0.1 m^3 and 0.3 m^3, and -0.3 m^3, respectively. Over (multiple) one day periods the upper braidplain and lower braidplain zones of -0.3

m^3 and 1.6 m^3, and -0.4 m^3, respectively. In the proglacial zone there was an order of magnitude difference in this crude measure of geomorphological activity between annual and daily time periods, which are characterised by net erosion and net deposition, respectively. In the upper braidplain there was a trend of net erosion on annual timescales and net deposition on daily timescales, whereas in the lower braidplain the reverse was observed; a trend of net deposition on annual timescales and net erosion on daily timescales.

Besides the mean values described above, Figure 8 also illustrates the maximum, minimum and thus the range of elevation changes for multiple time periods on terrain classified as active river. This graphic (Fig. 8) thus illustrates the variability in quantified geomorphological activity on different time scales. The data show that longer time periods have more variability, nor that winter months have less variability than summer months, nor that the proglacial zone is more variable; i.e. has a greater range of elevation changes; and thus a greater range of intensity of geomorphological activity, in any time period than the upper braidplain, which is more variable than the lower braidplain (Fig. 8).

The estimated net volume change (total deposition minus total erosion) for all substrate classes combined (except for the glacier and for known snow patches) over two years (Table 3) equalled 1875 m^3 for the lower braidplain, -1313 m^3 for the upper braidplain, and -4035 m^3 for the proglacial area. Across the whole area, there was a net loss of material of -5112 m^3 for 2009 and -3703 m^3 for 2010; a total of -8815 m^3 over the two year study period. This is a mean -4407 m^3 per year and hence we calculated a mean surface lowering rate distributed across the combined area of the three zones of interest (Fig. 2B) of -34.4 $mm.yr^{-1}$. If it is assumed that the upper part of the Ödenwinkelkees catchment was geomorphologically active during the study period, albeit to a lesser extent and intensity than the lower part, then this volume change provides a minimum estimate of mean surface lowering across the whole catchment of at least -4.8 $mm.yr^{-1}$. Volumetrically, the active river accounted for -4118 m^3 (71%) out of a total of -5826 m^3 material moved within the lower part of the Ödenwinkelkees catchment.

DISCUSSION: GEOMORPHOLOGICAL ACTIVITY AND DENUDATION

Sediment fluxes within alpine catchments and other cold environments are particularly affected by the effects of ice and snow on the landscape (e.g. Warburton, 1999, 2007; Slaymaker, 2010) and thus are highly sensitive to environmental change. Contemporary glacial processes, processes intrinsic to past glaciations, direct transport processes related to frozen water (avalanches, slush flows), ground ice dynamics and phase changes of water resulting in sediment mobilisation all exert a strong control on the spatial and temporal distribution and intensity of geomorphological activity and sediment fluxes (Beylich and Warburton, 2007). Overall, there is some quantitative data on these discrete processes in the literature but there is far less understanding of the nature of the links between them and of the variability of these links (e.g. Korup, 2002). The landsystem approach does not help this understanding because it cannot account for different timescales of evolution between components but also importantly it cannot consider connections between different components.

Sediment fluxes from alpine and other cold environments are dominated by proglacial fluvial processes (e.g. Hewitt, 2002) and research efforts into understanding sediment fluxes in alpine regions have concentrated on the fluvial part of the geomorphological landsystem (e.g. Bertoldi, et al. 2009). Our project is not any different in its (proglacial) focus and thus unsurprisingly our sediment flux data (Table 3), which pertain to two years of representative weather conditions (Fig. 4), show a dominance of the fluvial system in geomorphological activity throughout the proglacial area of the catchment. However, if adjacent cells with a similar elevation change can be considered as a ‘zone’ (perhaps as a landform), then adjacent ‘zones’ of erosion and deposition, and ‘zones’ that switch from erosion to deposition and vice versa in our data (Figs. 5, 6 and 7, Table 3) suggest that mass movement and alluvial/colluvial sources of sediment are often coupled to the fluvial system, albeit in an episodic fashion. A similar finding was reported by Schrott et al. (2006) for the deglaciated Reintal Valley in Germany. Consequent to such coupling, sediment transfer rates can either be extremely slow such as for solifluction or very fast such as for slope failures. We also note that over very short time-scales even fluvial sources of sediment in the Ödenwinkelkees catchment; i.e. glacial ice, seasonal snow and groundwater/permafrost sources, were all ephemeral and thus very variable both in time and in space. This variability clearly produced rapidly changing channel morphology and continual exploitation of new sediment sources; Figs. 5, 6 and 7, (c.f. Warburton, 1990; Hodgkins et al., 2003; Morche et al., 2008). The large lateral (LIA) moraines are the dominant feature surrounding the Odenwinkelkees glacier margin and these reflect glacier advance and thickening followed by stagnation and downwasting; rather than glacier retreat. These moraines are clearly temporary sediment storages and intermittent sediment sources (Fig. 7, Table 3). It is worthwhile noting that sediment stores within alpine catchments have previously been attributed to be formed primarily by gravitational and nival processes, but destroyed by fluvial processes (Schrott et al. 2006). In part, our contemporary data illustrate the inverse of these attributes; namely the erosion of hillslopes by gravitational processes (falls, slides, slumps) and the construction of landforms by fluvial processes (bar, bank and fan aggradation).

Contemporary geomorphological activity in the Ödenwinkelkees catchment does not occur homogenously in space; it is fragmented, although it does have an identifiable spatial pattern (Figs. 5, 6 and 7). This is perhaps not a surprise where terrain is classified as ‘active river’, but the characteristic extends to other parts of the catchment that are classified as alluvial/colluvial, boulders and bedrock. This ‘spatial fragmentation’ of geomorphological activity will be missed by studies solely concentrating on individual landforms. Further studies should look to quantify the spatial coherence/fragmentation of geomorphological activity. Furthermore, much of the geomorphological activity both within the river and on hillslopes has neither a point source, nor a clearly defined transport route (Figs. 5, 6 and 7). This means that studies based on within-catchment at-a-point hydrological monitoring programs could be under-estimating sediment flux considerably dependent simply upon study location.

For all substrate classes (except ‘glacier’) within the Ödenwinkelkees catchment we find a trend of increasing area, volume and rates of sediment movement with decreasing duration between surveys (Table 3). We accept that this trend could result because our shorter survey intervals were biased towards the summer months, but by examining the sequential DEMs of difference (Figs. 5, 6 and 7) we interpret this trend as strong indicator of the composition and behaviour of contemporary geomorphological activity; specifically that erosion events are followed rapidly by depositional events and vice versa. High frequency sediment transport events and geomorphological change within proglacial rivers have been quantified by Ferguson et al. (1992), Ashworth et al. (1992) and Lane et al. (1995), for example. Goff and Ashmore (1994) used video evidence to show that over the scale of several bar lengths channel change is frequent and involves the destruction and construction of bars, with sections of channel being intermittently abandoned and reoccupied. Sambrook-Smith (2000) examined high-frequency proglacial sedimentation and identified preservation potential and diagnostic characteristics; i.e. cyclicity with distance away from the glacier. This agrees with our results from the ‘lower braidplain’ zone where geomorphological events were not so intense but more frequent (Fig. 5, Table 3). Similarly, but with application of repeated TLS surveys, Milan et al. (2007) reported re-working of proglacial river sediments by bedload transport. However, our datasets (Figs. 5, 6 and 7), which are novel in being unrestricted to specific landforms and of multiple temporal intervals, demonstrate that infilling of recently excavated hollows and mobilisation of recently deposition sediments, whether by fluvial, alluvial/colluvial or mass movement processes are common across the catchment. The reworking of recently mobilised sediments, and the frequency of this geomorphological activity, is herein found to be a lot more widespread and intense across alpine catchments than previously measured.

Denudation rates estimated from contemporary processes need to be reconciled with those based on longer-term (centennial-millennial) time-scales (Kirchner et al., 2001) Our calculated contemporary mean surface lowering rate of at least 4.8 $mm.yr^{-1}$ for the proglacial part of the Ödenwinkelkees catchment is ~ 2.5 times that of 2 $mm.yr^{-1}$ determined for; i) the last 6000 years in study of alpine catchments in the Himalaya by Shroder et al. (1999), ii) late Holocene rates from British Columbia (Owens and Slaymaker, 1993). It is also double the values of 0.1 – 2.6 $mm.yr^{-1}$ and 0.2 – 1.4 $mm.yr^{-1}$ most recently reported from the European Alps by Otto et al. (2009) and by Norton et al. (2011), respectively. Our rate is higher than previous estimates perhaps because it pertains only to the proglacial part of the catchment, perhaps because it is of unconsolidated sediment rather than bedrock, and perhaps because it is a shorter (contemporary) time period of study. Additionally, and interestingly, it encompasses geomorphological activity across several types of (substrate type) terrain within a catchment, rather than being restricted to just sediment exiting a catchment due to fluvial processes. We would suggest qualitatively that surface lowering takes place in sediments that are dominantly provided by the retreating glacier. The erosion of this glacial sediment is probably decoupled from the supply of this sediment. Therefore if glacial retreat rates and land surface ages could be obtained, such as from dated moraines (Fig. 2A) future studies could seek to obtain a quantitative link between sediment supply and land surface age; i.e. deglaciation.

There is an important need to reconcile contemporary sediment transfer rates estimated from repeated geomorphological surveys with those from hydrological monitoring (Warburton, 1990). Our rate of 4.8 $mm.yr^{-1}$ for the whole Ödenwinkelkees catchment as obtained from repeated surveys is comparable to the values of up to 1 - 10 $mm.yr^{-1}$ presented by Hallet et al. (1996) from hydrological monitoring studies and sedimentation rate measurements on mountain catchments in Alaska, British Columbia and the Himalaya. It is rather less than the regional denudation rate in New Zealand calculated to be 9 $mm.yr^{-1}$ due to a volumetric analysis of landsliding by Hovius et al. (1997), but very similar to the values reported in the synthesis of contemporary mountain denudation rates by Hicks et al. (1990) from sedimentation in New Zealand lakes. Whilst our rate of 4.8 $mm.yr^{-1}$ includes all types of geomorphological processes, it is interesting to note that Riihimaki et al., (2005) were able to deconstruct the sediment flux from an Alaskan catchment to show the effectiveness and dominance of mountain glacier processes in lowering bed elevations by 1 - 2 mm.yr^{-1}.

Hallet et al. (1996) showed a high geographical variability in denudation rates but noted a pattern of increased denudation rates with total catchment area, with the volume of ice within a catchment and with the discharge of water from a the catchment. However, there are a number of circumstances where such relationships will not apply (Owens and Slaymaker, 1992). Hicks et al. (1990) drew attention to the contrasts in glaciated versus non-glaciated catchments and Harbor and Warburton (1992, 1993) advocated the relative importance of sediment storage. We also found that temporary sediment storage is profoundly important. This is the reason why our mean denudation rate is relatively high; because we measured total geomorphological activity, which includes intra-catchment mobilisation and temporary storage, rather than a basin-averaged efflux. It is also the reason why we found that 71% by volume of measured geomorphological activity within the Ödenwinkelkees catchment over a (relatively short) two-year time-span was due to fluvial processes that operated across an area < 5% of the total catchment.

CONCLUSIONS AND WIDER IMPLICATIONS

Alpine catchments are especially sensitive to climate change and to human impacts. It is imperative to understand contemporary geomorphological activity within alpine catchments in order to separate these impacts accurately from natural weathering rates. However, a holistic discrimination of sediment sources, quantification of sediment fluxes, characterisation of geomorphological activity by substrate class and the inter- and intra-annual spatial and temporal variability in these has hitherto been unreported. High-resolution changes to individual landforms have been measured on an episodic campaign basis, but this technique has not been applied to entire proglacial areas. Quantification of geomorphological work within alpine catchments has been restricted to hydrological gauging of total suspended sediment.

This study made a novel discrimination and quantification in space and time of contemporary multi-scale geomorphological changes within the proglacial part of a glaciated alpine catchment; the Ödenwinkelkees, Austria. This permitted novel quantitative measurements of the relative importance of contemporary geomorphological activity on different substrate classes to total sediment flux / denudation and also that of episodic versus continuous processes. However, we could not measure all sediment sources; sediment input from the glacier or from rock falls in the upper catchment, such as have been documented in the neighbouring Pasterze catchment (e.g. Kellerer-Pirklbauer et al., 2012) were not captured due to the range of the TLS and due to the accessibility of suitable vantage points. Repeated surveying permitted calculation of a net volume of material moved of 4400 m^3 per year in the lower part of the Ödenwinkelkees catchment. This equalled 34 $mm.yr^{-1}$ for this part of the catchment. If it is assumed that the upper part of the catchment at least has some geomorphological activity surface lowering across whole catchment was at least 4.8 $mm.yr^{-1}$. Repeated surveying therefore appears to mitigate the problem identified by Kirchner et al. (2001) of hydrological methods underestimating sediment fluxes on the annual – decadal scale.

The net volume change for all substrate classes combined (except for the glacier and for known snow patches) over two years equalled 1875 m^3 for the lower braidplain, -1313 m^3 for the upper braidplain, and -4035 m^3 for the proglacial area. The upper part of the proglacial area is therefore in net degradation / erosion whilst the lower part is in net aggradation / deposition (Fig. 9). Geomorphological activity decreased in both areal density (different substrate classes and number of pixels) and in intensity (magnitude of elevation changes) from the proglacial zone to the upper braidplain zone to the lower braidplain zone, i.e. with distance from glacier (Fig. 9); this is consistent with the paraglacial concept. In the fluvial substrate class this relationship with distance from glacier was not smooth (Fig. 9) because moraines form a local topographic constriction and a locally elevated base level. Overall, hillslope activity dominated in the mid-sections of the study area and fluvial activity became progressively more important (areally and volumetrically) towards the lower part of the study area. The summary model in Figure 9 therefore not only represents the spatio-temporal geomorphological activity, but thereby infers terrain stability and the likely preservation or persistence of landforms and sediments. We note that 71% of geomorphological activity by volume was restricted to < 5% of the catchment area.

Temporally, there was a concentration of geomorphological activity in the Spring and Summer months, which highlighted the importance of phase changes of water in alpine catchments for sediment fluxes. Up to a period of two years, a longer duration of study produced more variability in active area and more variability in the magnitude of elevation changes as geomorphological activity occurred sporadically. Within this time period there was a dominance of continuous processes over episodic processes.

More widely, this study provides a method by which inter-annual and inter-catchment variation (Lenzi et al., 2003; Carrivick and Rushmer, 2009; Carrivick and Chase, 2011) can be quantified, and by which comparison of different catchments (Gurnell et al., 1988; Beylich et al., 2006; Warburton et al., 2007; Carrivick and Rushmer, 2009; Carrivick and Chase, 2011) can be made. Future studies should look to i) utilise repeated ALS to determine geomorphic changes over a whole catchment, ii) quantify the spatial organisation/fragmentation of geomorphological activity, and iii) quantify sediment supply with land surface age; i.e. with deglaciation. With respect to these latter two topics, this study provides a conceptual framework by which the contemporary importance of glacial processes versus non-glacial processes can be measured (Hicks et al., 1990; Harbor and Warburton, 1992, 1993; Hallet et al., 1996), and thus changes in these processes due to climate (Hodgkins et al., 2003; Stott and Mount, 2007) and human impacts can be predicted.

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Zone of interest and in parentheses area in km² (for location see Figure 2B)	Number of single fine scans and in parentheses number of surface planes used (valid plane defined with 5 m max. edge and 0.2 m plane error)	MSA results or 'Error' (m) (as standard deviation in m of normal distance between the used planes)
Lower braidplain (0.07)	9 (17,536)	0.0324
Upper braidplain (0.15)	14 (81,050)	0.0334
Proglacial (1.05)	14 (820,585)	0.0369

Table 1: Number of single fine scans and plane surfaces used in the Multi Station Adjustment and overall errors as standard deviation

Zone	Survey date	Grid cell size (m)	Points	NO points	Coverage
Lower braidplain (0.07 km^2)	20/07/2008	0.5	289,300	2,500	99.14%
	28/06/2010	0.2	142,134	1,668,682	7.85%
		0.5	104,257	186,120	35.90%
		1.0	47,001	25,863	64.51%
	29/06/2010	0.2	885,714	925,102	48.91%
		0.5	222,605	67,772	76.66%
		1.0	64,374	8,490	88.35%
	31.06.2010	0.2	170,937	1,639,879	9.44%
		0.5	114,140	176,237	39.31%
		1.0	49,205	23,659	67.53%
	25/08/2010	0.2	323,280	1,487,536	17.85%
		0.5	154,412	135,965	53.18%
		1.0	53,129	19,735	72.92%
Upper braidplain (0.15 km^2)	02/07/2008	0.2	140,541	1,310,994	9.68%
		0.5	140,541	92,276	60.37%
		1.0	58,080	5,437	91.44%
	20/07/2008	0.5	633,420	2,968	99.53%
	26/08/2008	0.2	249549	1,903,848	11.59%
		0.5	249549	95,152	72.40%
		1.0	76319	9,835	88.58%
	29/08/2008	0.2	269,371	2,535,500	9.60%
		0.5	269,371	180,236	59.91%
		1.0	92,393	20,091	82.14%
	28/06/2010	0.2	1,778,000	2,178,601	44.94%
		0.5	470,217	163,824	74.16%
		1.0	128,910	30,011	81.12%
	29/06/2010	0.2	1,584,905	2,371,696	40.06%
		0.5	479,386	154,655	75.61%
		1.0	144,256	14,665	90.77%
	30/06/2010	0.2	606,764	3,349,837	15.34%
		0.5	227,999	406,042	35.96%
		1.0	85,170	73,751	53.59%
	31.06.2010	0.2	1,926,863	2,029,738	48.70%
		0.5	535,930	98,111	84.53%
		1.0	148,395	10,526	93.38%
	25/08/2010	0.2	1,954,452	2,002,149	49.40%
		0.5	533,503	100,538	84.14%
		1.0	148,826	10,095	93.65%
Proglacial (1.05 km^2)	20/07/2008	0.5	8,033,820	4,766	99.94%
	25/08/2009	0.2	4,560,366	21,299,475	17.63%
		0.5	1,666,863	2,471,218	40.28%
		1.0	611,034	423,422	59.07%
	28/06/2010	0.2	1,465,334	24,645,209	5.61%
		0.5	857,620	3,320,207	20.53%
		1.0	373,981	670,534	35.80%
	31.06.2010	0.2	6,207,487	19,925,692	23.75%
		0.5	2,249,697	1,931,699	53.80%
		1.0	724,702	320,686	69.32%
	25/08/2010	0.2	2,404,100	23,780,744	9.18%
		0.5	1,217,416	2,972,227	29.06%
		1.0	499,793	547,651	47.72%

Table 2. List of airborne and terrestrial survey laser scan data and classification of coverage by number of survey points, per 0.2, 0.5 and 1.0 m grid cell

Substrate class (% area for each zone)	Time-frame	Lower braidplain area (m^2)	Lower braidplain Volume (m^3)	Lower braidplain Rate (m^3day^{-1})	Upper braidplain Area (m^2)	Upper braidplain Volume (m^3)	Upper braidplain Rate (m^3day^{-1})	Proglacial Area (m^2)	Proglacial Volume (m^3)	Proglacial Rate (m^3day^{-1})
River (L = 32%; U = 36%, P = 8%)	two years	18,323 (81%)	325	0.5	50,708 (94%)	-505	-0.7	73,530 (91%)	-3,938	-4.5
	one year		N/A			N/A		35,273 (44%) 66,988 (83%)	-869 -2,006	-2.3 -6.1
	two months	16,719 (74%)	88	2		N/A		41,288 (51%)	-627	-8.0
	3 days	17,114 (76%)	-70	-23	49,984 (92%) 41,319 (76%) 32,344 (60%)	-173 13 -35	-58 4 -4	51,240 (63%)	-64	-23
	one day	18,114 (80%)	-18	-18	32,275 (60%) 49,114 (91%)	61 -16	61 -16		N/A	
Boulders (L = 3%, U = 25%, P = 10%)	two years	1,770 (89%)	64	0	6,052 (16%)	-605	-1		N/A	
	one year		N/A			N/A		44,654 (41%) 85,046 (79%)	623 -6	2 -3
	two months	1,746 (87%)	66	0	17,207 (45%) 5,669 (15%)	-562 -862	-10 -15	54,269 (50%)	148	2.5
	3 days	1,746 (87%)	-3	-1	17,084 (45%) 5,557 (15%)	-30 91	-10 30	58,057 (54%)	-107	-36
	one day	1,717 (86%)	93	93	30,938* 30,758* 16,713*	-202* -512* 351*	-202* -512* 351*		N/A	
Bedrock (L = 57%, U = 8%, P = 24%)	two years	129 (0.3%)	-11	0	1,682 (14%)	-68	0	19,040 (7%)	-97	-0.1
	one year		N/A		3,885 (33%) 1,576 (13%)	-109 -123	-0.2 -0.3	8870 (3%) 2151 (8%)	2,672 -1608	7 -4
	two months	124 (0.3%)	10	0		N/A		8548 (4%)	170	3*
	3 days	122 (0.3%)	-3	-1	1,011 (9%) 3,687 (31%)	0 -556*	0 -185*	7460 (3%)	-25	-8*
	one day	123 (0.3%)	-1	-1	914 (8%) 1059 (9%) 379 (3%)	9 -2 8	1 -2 8		N/A	
alluvial/colluvial (L = 9%, U = 31%, P = 46%)	two years	5,104 (85%)	1,497	2	5,355 (10%)	-135	0	405,244*	-2,348*	-3*
	one year	4,312 (72%)	297	0	5,255 (10%) 2,456 (4%)	20 18	0 0	355,346*	-1,640*	-5*
	two months		N/A			N/A		230,226*	2,845*	47*
	3 days	4,312	-142	-47	3,593	-14	-5	246,434*	-3,139*	-1046*

		(72%)			(7%)					
					5,530 (10%)	-10	-3			
	one day	4,301 (72%)	-97	-97	3,956 7%)	-21	-21		N/A	
					3,952 (7%)	56	56			
					5,419 (10%)	-49	-49			
Glacier (-,-,P = 11%)	two years		N/A			N/A		39,041	-174,229	-230
	one year		N/A			N/A		4,396	-8,590	-28
								57,761	-218,689	-558
	two months		N/A			N/A		4,024	-5,236	-90
	3 days		N/A			N/A		3,246	-1,216	-405

Table 3: Deposition (positive values) and erosion (negative values) by substrate class and by time interval between surveys. Values in parentheses are the % area of active cells within each substrate area, by zone of interest. Asterisks denote uncertain values due to probable snow patch melt.

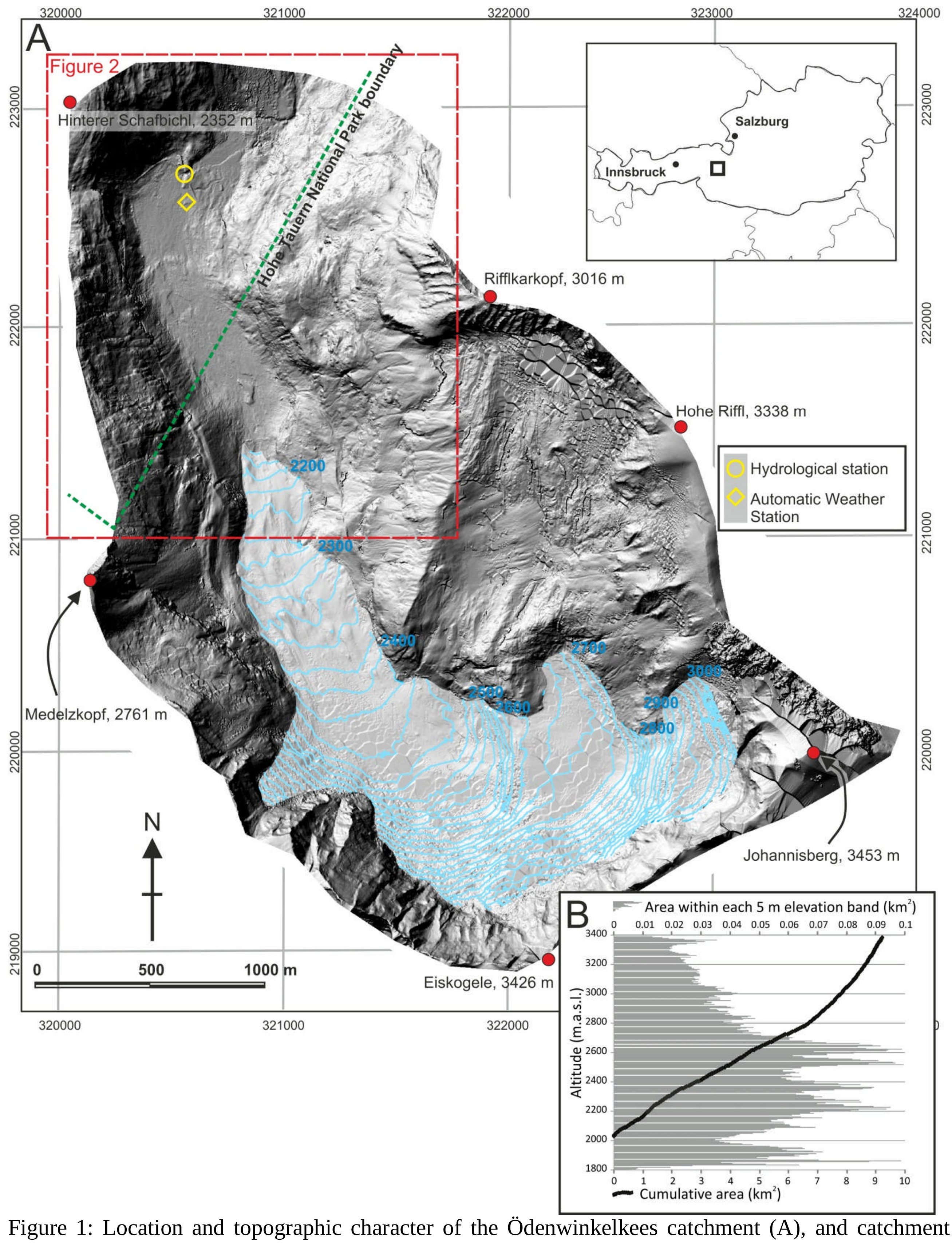

Figure 1: Location and topographic character of the Ödenwinkelkees catchment (A), and catchment hypsometry (B).

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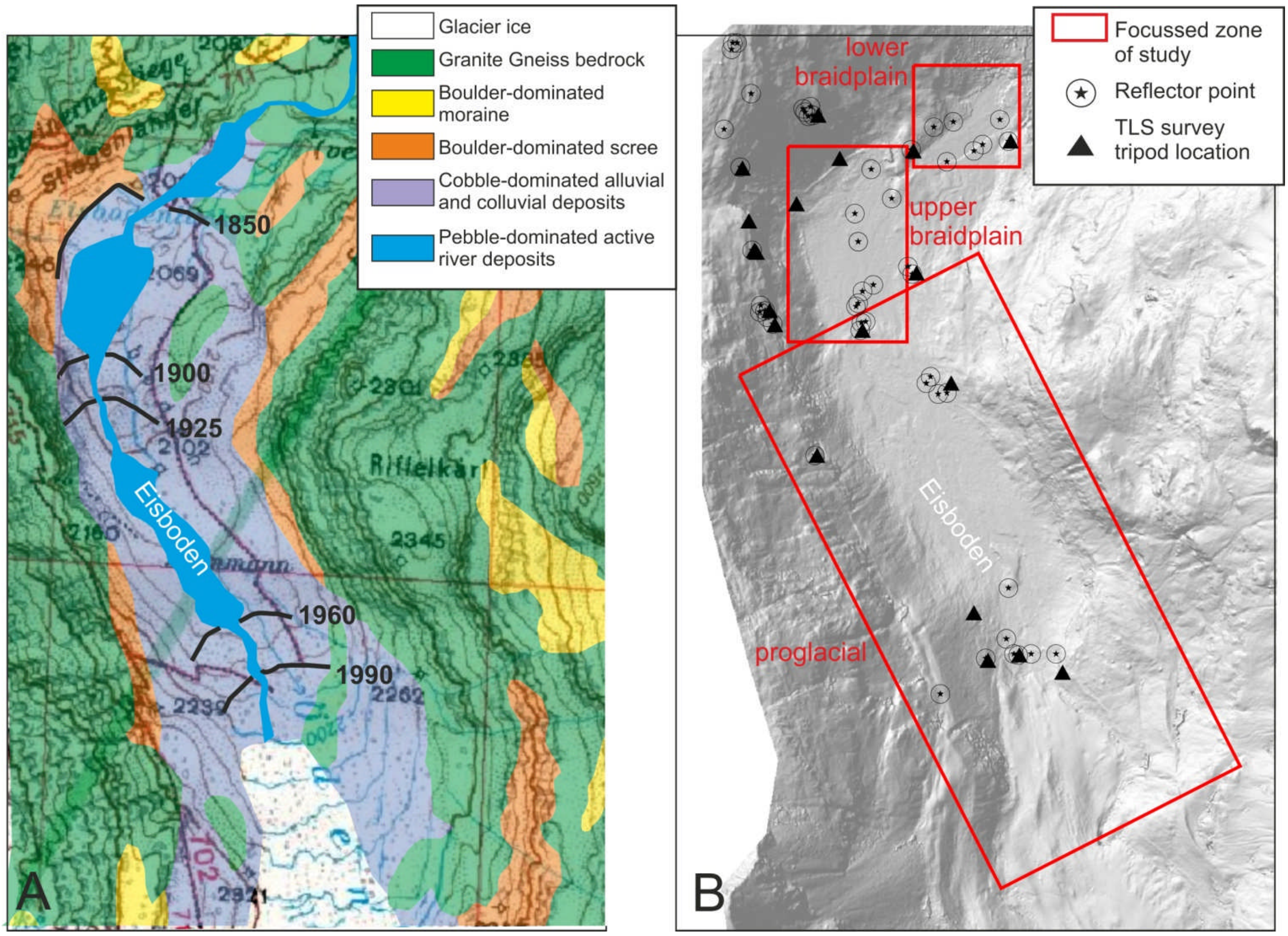

Figure 2: Northern part of the Ödenwinkelkees catchment substrate classification with dates of prominent moraines (black arc lines) adapted from Slupetsky and Teufl (1991) (A), and survey design (B)

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Figure 3: Ödenwinkelkees catchment in July 2008 illustrating topography, geomorphology and substrate. The elevation range of the glacier is ~ 800 m.

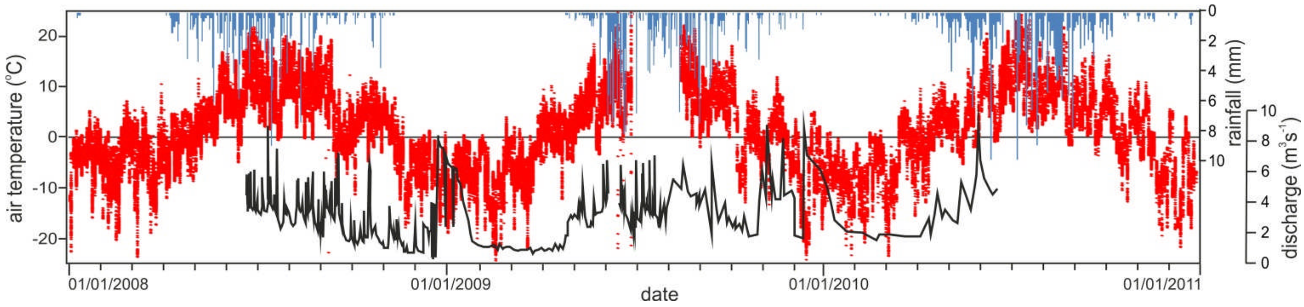

Figure 4: Weather conditions and river discharge during the study period recorded at the Automatic Weather Station and Hydrological station locations (Fig. 1), respectively.

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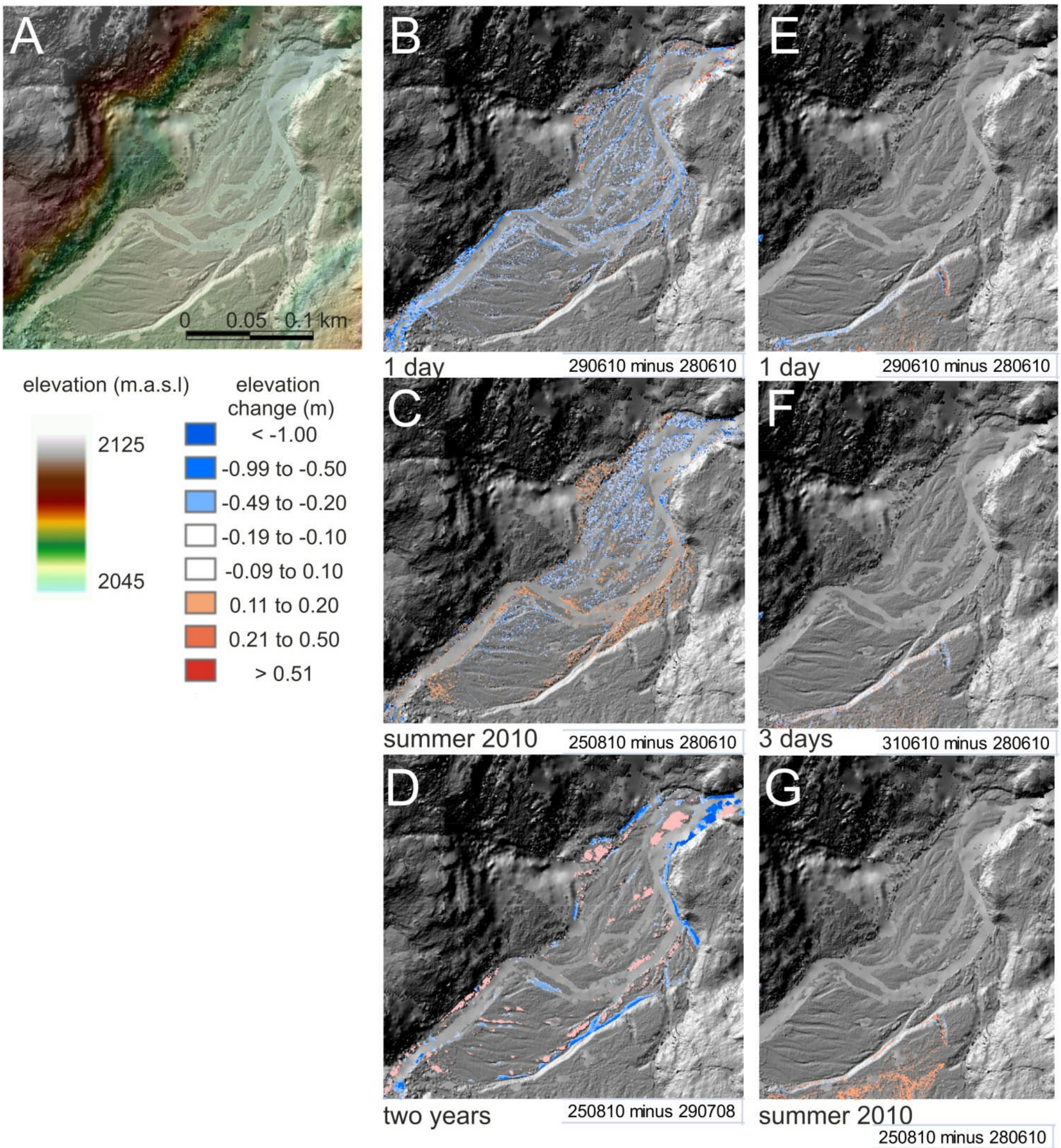

Figure 5: Spatial distribution and temporal intensity of surface elevation changes in the 'lower braidplain' zone. Panels B, C and D are changes on terrain classified as 'active river', and panels E, F and G are changes on terrain classified as alluvial/colluvial.

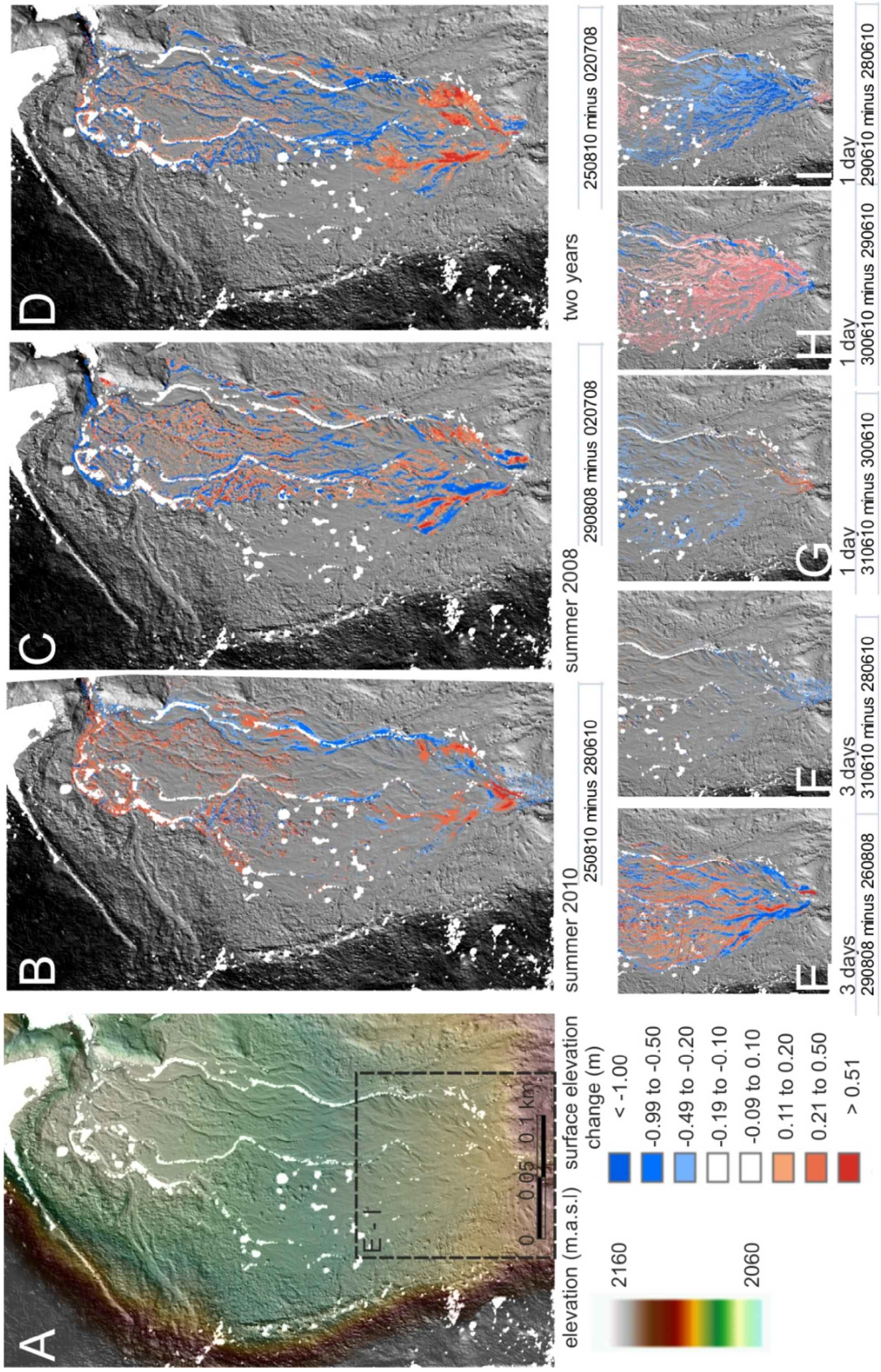

Figure 6: Spatial distribution and temporal intensity of surface elevation changes in the 'upper braidplain' zone. Panels B, C and D are changes on terrain classified as 'active river', and panels E - I are focussing on the fan apex.

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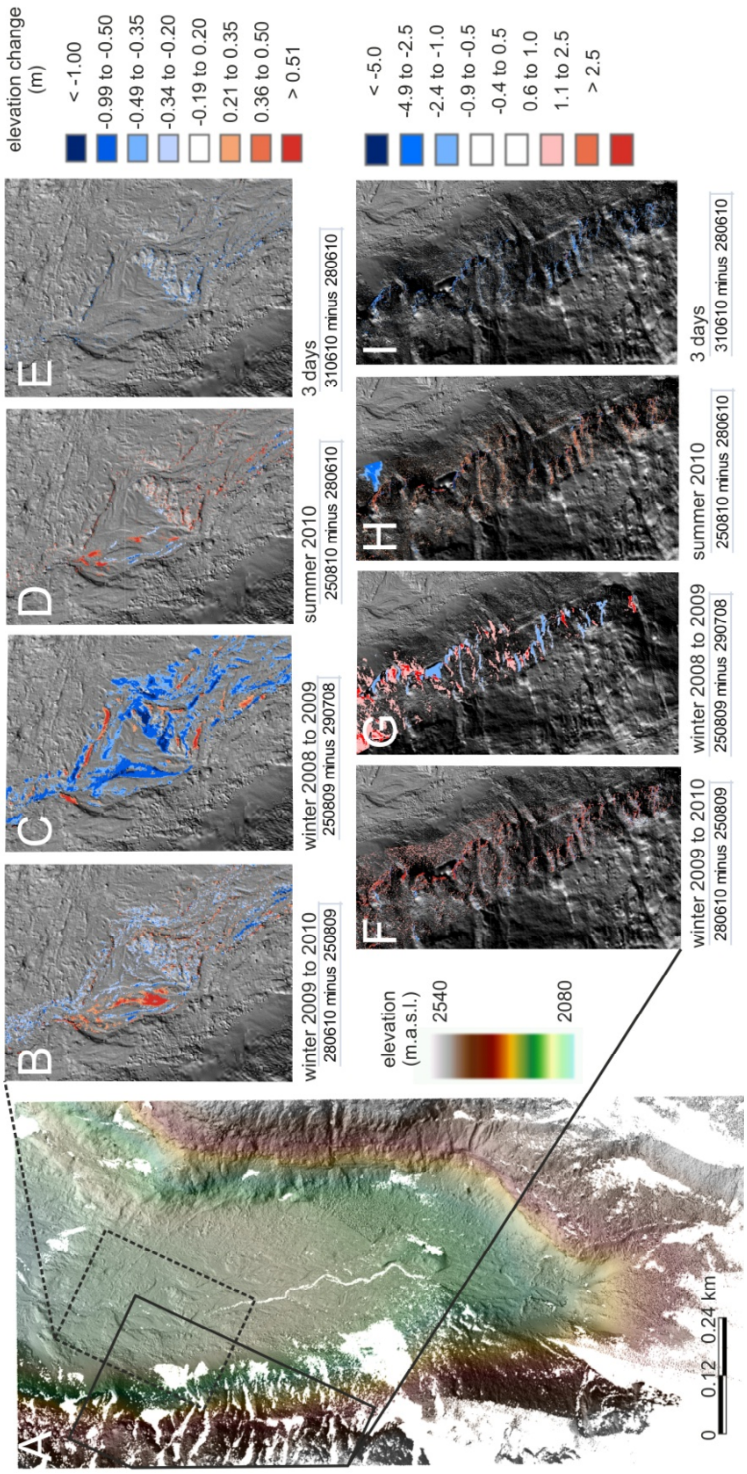

Figure 7: Spatial distribution and temporal intensity of surface elevation changes in the 'proglacial' zone. Panels B - E are changes on part of the terrain classified as 'active river', and panels F - I are changes on terrain classified as 'alluvial/colluvial'.

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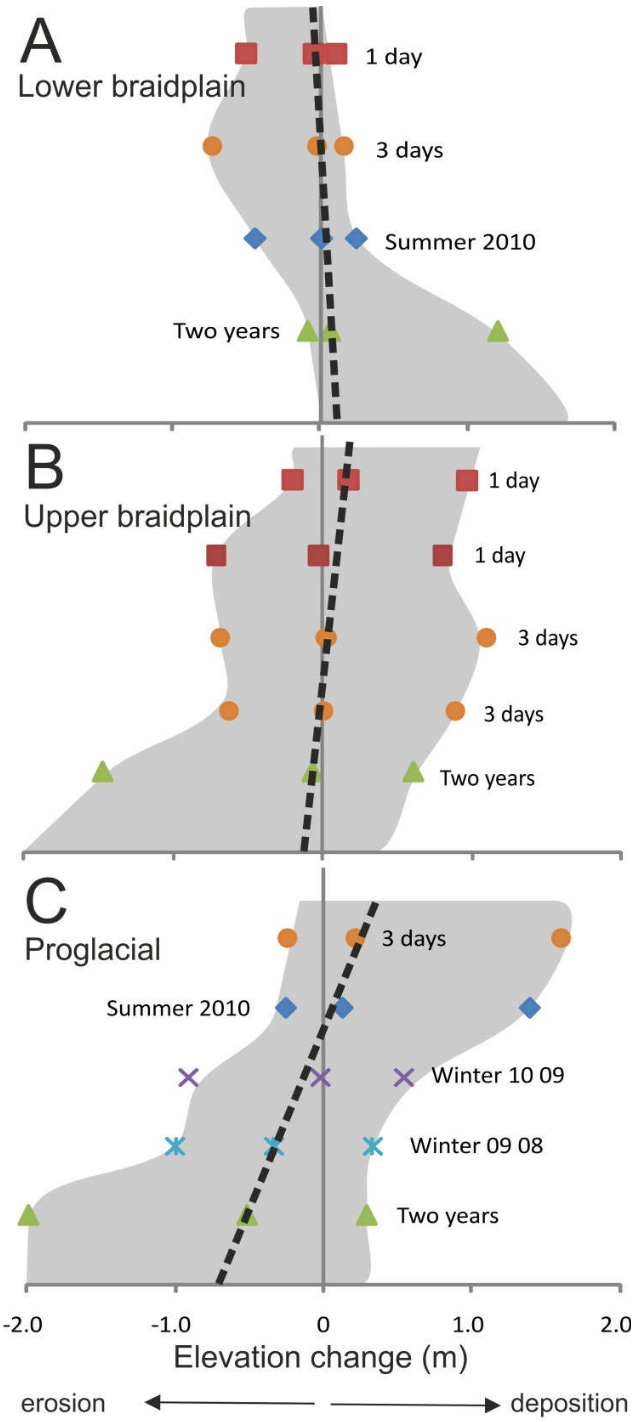

Figure 8: Variability of surface elevation changes with duration between surveys for terrain classed as ‘active river’ in the lower braidplain (A), upper braidplain (B) and proglacial (C) zones.

Contemporary geomorphological activity throughout the proglacial area of an alpine catchment

Carrivick et al.,

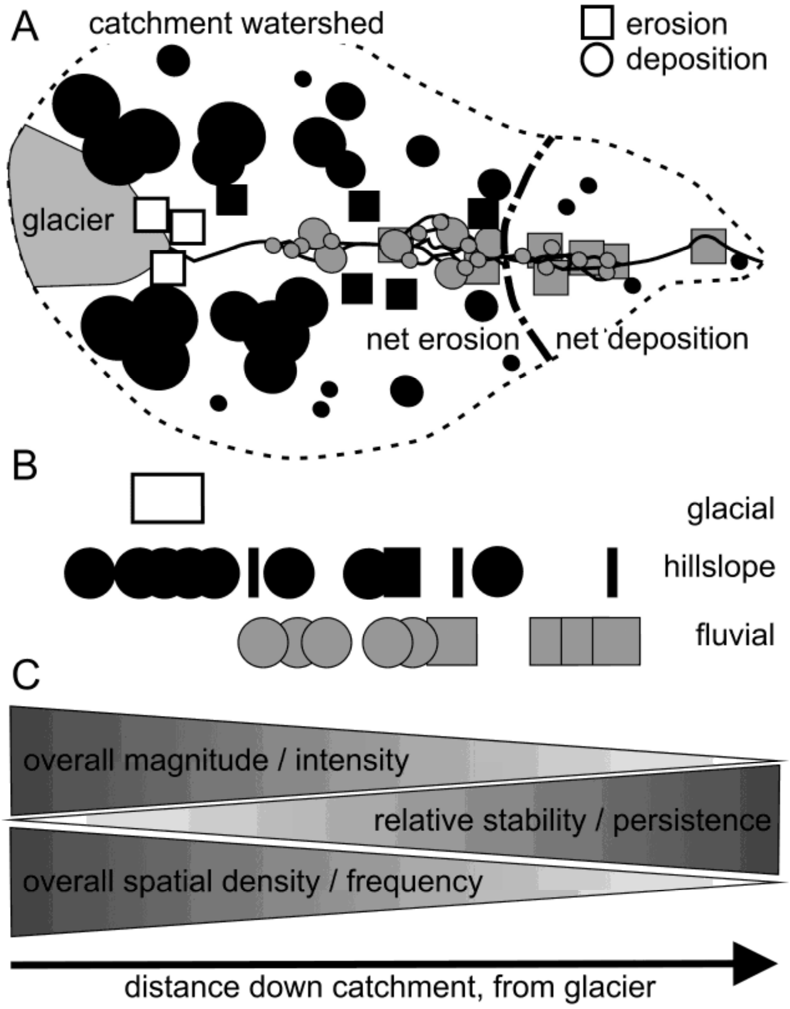

Figure 9. Summary conceptual model. A: Spatial variability in elevation changes (geomorphological activity) within the proglacial area of an alpine catchment. Symbols discriminate type of geomorphological activity; circles = erosion, squares = deposition. Symbol size denotes relative spatial intensity and spatial density of activity, and circle colour refers to categories of geomorphological processes; white = glacial, black = hillslope and grey = fluvial. B: Longitudinal trend in activity discriminated by process types; note the dominance of hillslope activity in the mid-sections of the catchment and the dominance of fluvial activity in the lower part of the catchment. C: Summary qualitative longitudinal pattern of geomorphological activity. Note that for clarity the quantitative nature of the measurements made within this study are not represented, and that the category of ‘hillslope’ processes include those on ‘bedrock’, ‘boulders’ and ‘alluvial-colluvial’ substrate.