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Changes in sediment connectivity following glacial debuttressing in an Alpine valley system

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Abstract

Increasing air temperature and declining winter snowfalls are resulting in rapid glacier recession and the expansion of proglacial margins in Alpine regions. Such margins include substantial debris accumulations (e.g. frontal/lateral moraine ridges; till-covered and steep valley sidewalls) which may be unstable due to glacial debuttressing. Rainfall, snowmelt and ice melt out may then cause mass movements. Here, we quantify the decadal-scale erosion and deposition patterns and changes in connectivity for two valley sidewall geomorphological systems following retreat of the Glacier d'Otemma, Switzerland. We apply archival digital photogrammetric methods to the period 1964 to 2009 to determine high resolution digital elevation models. These were differenced to calculate patterns of erosion and deposition and to quantify the evolution of sediment connectivity. We found that gully headward erosion (rates between ca. -10.6 mm/yr and -1002.1 mm/year) was the main geomorphological process during glacier thinning but increasing depositional rates downslope of the gullies (ca. +21.3 to +298.5 mm/yr) were recorded in the following years associated with significant alluvial fan growth at the slope base. Whilst gullying enhanced connectivity by removing glacially conditioned sediment transfer buffers, so connecting side-slopes to upstream sediment sinks (the upslope contributing area between 1964 and 2009 increased by +73.8% and +195.1% in each subsystem), alluvial fans reduced the rates of sediment transfer to the rapidly enlarging glacial forefields. The detail of these responses is conditioned by three generic processes: (1) the wider geomorphic setting – here, the presence of a moraine bastion as a primary part of the sediment cascade strongly influenced gully morphology evolution and the likely length of the paraglacial period length; (2) the thickness of sediment left by the retreating glacier which controlled the influence of bedrock topographic buffers on connectivity; and (3) the extent to which diffusive drainage systems develop in response to the deposition at the hillslope base, which tends to disconnect sediment flux. Post-glacially, gully development has a self-limiting effect on sediment connectivity in that whilst gullying increases sediment connectivity, the eroded sediment leads to deposition on the alluvial plain that reduces sediment connectivity.

Keywords: Climate warming; Glacial debuttressing; High mountain regions dynamics; Paraglacial; Sediment connectivity; Historical evolution.

Highlights:

- Presents one of the first multi-decadal scale records concerning valley sidewalls response, in term of geomorphological processes and connectivity, to rapid glacial retreat and debuttressing.
- Connectivity develops in response to glacial recession and thinning by gully headward erosion.
- Connectivity impacts morphodynamics, and vice-versa. Paraglacial geomorphological features may extent the paraglacial period through secondary reworking of sediment.
- Gullying process increase the vertical connectivity (basin scale) but it may decrease the lateral connection within the gully network (local scale).

1. Introduction

High mountain regions are strongly sensitive to climate change. The rapid recession of Alpine glaciers and permafrost zones following recent atmospheric temperature rise is well documented (Haeberli et al., 1997; Haeberli and Beniston, 1998; Fischer et al. 2014, 2015). Since the end of the Little Ice Age (LIA), mean annual atmospheric temperatures (MAATs) in the European Alps have increased by about +2°C, generating widespread negative glacier mass balance and ice volume loss (Bauder et al., 2007; Fischer et al., 2015). European Alpine glaciers lost c. 50% of their surface between 1850 and 2000 while Swiss glaciers, based on the last Swiss glacier inventory of 2010, retreated by c. 40% (Paul et al., 2004; Zemp et al., 2006; Fischer et al., 2014). The result has been a substantial increase in the size of proglacial margins, by 920 km^2 in Switzerland and Austria since the end of the LIA (Carrivick et al., 2018).

Proglacial margins are characterized by over-steepened slopes and unconsolidated sediment accumulations, often unstable, and acting as potential sediment sources (Ballantyne, 2002a,b; Laute and Beylich, 2014) notably during extreme events (Wulf et al., 2012). The period of landscape reworking that follows is commonly referred to as “paraglacial”, ending with a new and more stable deglaciated state (Church and Ryder, 1972; Cossart et al., 2008; Porter et al., 2010; Carrivick and Heckman, 2017). During the paraglacial phase, an increase in the availability of glaciogenic deposits may accelerate the down valley sediment cascade (Ballantyne, 2002a; Cossart and Fort, 2008) especially because such deposits are commonly poorly sorted, including a relatively easily eroded fine sediment fraction (Derose et al., 1998). The geomorphic consequences of glacier retreat are well described in the scientific literature. On over-steepened rock slopes, glacier debuttressing may result in major rockfalls and landslides as glacier retreat destabilizes slopes through undercutting and, at the same time, the progressive loss of pressure exerted by ice on slopes may result in rock stress-releases (Cossart et al., 2008; McColl, 2012; Davis et al., 2013; McColl and Davis, 2013). Moraine accumulations may act as important sediment sources after glacial retreat as shown for the European Alps (e.g. Curry et al., 2006; Cossart and Fort, 2008; Carrivick et al., 2013; Lane et al., 2017), Western Norway (e.g Ballantyne and Benn, 1994; Mercier et al., 2009; Laute and Beylich, 2012), and Canadian Rocky Mountains (e.g Hugenholtz et al., 2008). It is thought that the glacier advance that took place during the Little Ice Age slightly enhanced geomorphological activity during the subsequent paraglacial phase (Laute and Beylich, 2013), in terms of headward erosion of gullies and occurrence of debris flows, affecting debris covered slopes located inside Little Ice Age margins (Laute and Beylich, 2012).

On moraine accumulations, buried ice melt, sediment mass failures and erosion by water, including formation of debris flows, may encourage an erosional response (Evans and Clague, 1994; Laute and Beylich, 2012, 2014) and the formation of complex gully systems (Ballantyne and Benn, 1994; Lane et al., 2017). Hydrological processes may erode lateral moraines at measured rates of 49 to 151 mm/year in the European Alps (Curry et al., 2006), 50 to 100 mm/year in Norwegian glacierized catchments (Ballantyne and Benn, 1994) and 0.4 and 31 mm/year in the Nepalese Himalaya (Watanabe et al., 1998). Gully development acts as a source of sediment and encourages upstream to downstream connection of sediment but also lead to debris fan and alluvial fan formation at the slope base (Lane et al., 2017). Carrivick et al. (2013) reported alluvial fan growth in the Austrian Alps at a rate of 7 m^3/day due to contemporary reworking of moraine-derived material formed during the Little Ice Age. The diffusive nature of alluvial fans may reduce rates of sediment flux to the alluvial plain where it can be reworked fluvially, representing an important potential negative feedback towards the end of the paraglacial period (Church and Ryder, 1972; Mercier, 1997, 2008; Ballantyne, 2002a,b; Lane et al., 2017). Such changes may have wider impacts upon proglacial sediment budgets (Carrivick et al., 2013; Staines et al., 2015).

This body of work suggests that evolving sediment connectivity may provide a conceptual basis for explaining changing sediment flux during the paraglacial phase following rapid glacier recession (Baewert and Morche, 2014; Heckmann et al., 2016). Such understanding is crucial as the duration and intensity of the paraglacial phase will be partly a result of the balance between connectivity increasing (i.e. positive feedbacks) and connectivity-reducing (i.e. negative feedbacks) processes following glacier recession. The sensitivity of sediment flux to connectivity in paraglacial environments is likely to be increased because glaciers create landforms (e.g. lateral moraine ridges) which may interrupt the sediment cascade (Brardinoni and Hassan, 2006; Cossart and Fort, 2008; Fryirs, 2013; Lane et al., 2017). Such interruptions are unlikely to be permanent, as glacier retreat and thinning leads to base level fall that can activate gullying and headward erosion (Schiefer and Gilbert, 2007; Cossart and Fort, 2008) eventually dissecting such buffers (e.g. Gomez and Purdie, 2018; Lane et al., 2017).

This paper builds on past studies of proglacial landscape evolution following glacier retreat whether descriptive (e.g. Mercier, 1997; Ballantyne, 2002b), or more quantitative (e.g. Warburton, 1990; Ballantyne and Benn, 1994; Schiefer and Gilbert, 2007). Research has considered connectivity in deglaciating environments in general (e.g. Schrott et al., 2006; Carrivick et al., 2013; Baewert and Morche, 2014; Carrivick and Heckmann, 2017), and hydrological connectivity in particular (e.g. Cavalli et al., 2013; Lane et al., 2017). Here we aim to provide one of the first quantitative assessments of how hillslope sediment connectivity evolves at the decadal time-scale following from glacier thinning and retreat. Such research is not straightforward because unlike glaciers themselves, the hillslopes that form following their retreat are rarely monitored (Carrivick et al., 2015). Extracting topographic information from archival imagery using photogrammetry is one solution to this information and this has now been tested and applied for such zones over decadal time-scales (e.g. Schiefer and Gilbert, 2007; Micheletti et al., 2015a, 2015b; Staines et al., 2015).

The development of the software used in Structure from Motion Multi-View Stereo (SfM-MVS) photogrammetry (Westoby et al., 2012; Fonstad et al., 2013) opens up the potential of this kind analysis to a wider community even if the way it is used has to be modified for the large format imagery typical of archives (Bakker and Lane, 2017). The work is focused on a Little Ice Age lateral moraine in the Otemma Valley, south-west Switzerland, in a context of rapid climate warming and rapid glacier recession. It is representative of many Alpine valley glacier systems where glacial erosion has created a relatively wide valley floor combined with steep valley sides.

2. Methodology

2.1 Study site and climatic context

The proglacial margin of the Glacier d'Otemma is at an altitude of 2450 m a.s.l., just below the lower limit of the discontinuous permafrost belt (Lambiel and Reynard, 2001; Deluigi et al., 2017) (Figure 1). Mean Annual Air Temperature (MAAT) increased by +1.7°C during the 20th Century, but this increase was not uniformly distributed and there have been four main phases; a relative stable and cold period until the 1940s; a rapid warming lasting for about two decades; a cooler epoch between the 1960s and mid-1980s; and finally a warmer phase which extends until today. Since the 1960s, the start of our study, the increase is c. +1.4°C (Figure 2).

Even if this climatic trend is a common across the European Alps, evolution of the Glacier d'Otemma is different to most other glaciers in the region. With some exceptions (e.g. Gabbud et al., 2015), most Swiss Alpine valley glaciers saw a temporary re-advance in the 1970s and 1980s and then very rapid recession since (Haeberli and Beniston, 1998; Bauder et al., 2007, Fischer et al., 2015). It is only the longer and less steep glaciers that tend to show continual recession. The snout of the Otemma glacier retreated continually by 462 m between 1900 and 1967 (6 m per year, Figure 2). There was a temporary acceleration in retreat between 1967 and 1975 (to 42 m per year, Figure 2) suggesting some delay in glacier recession in response to the warming that started in the 1940s and reflecting the long length of the glacier (c. 9 km in 1964) and so the longer glacier response time. Recession of the Otemma glacier slowed from 1975 until the early 1980s (to 11 m per year; Figure 2), probably reflecting the relatively cooler period from the mid-1960s to the early 1980s. From the early 1980s, glacier recession rates increased to c. 32 m per year until the early 2000s and then to c. 50 m per year until present. The recession rate increases are probably explained by the reduction of winter snow accumulation by c. -50% in the region between the mid-1980s and 2010 (Micheletti et al., 2015b), coupled to higher MAATs. In summary, the Glacier d'Otemma has retreated by 2110 m since the 1950s, losing about 60% of its volume and 40% of its surface (Figure 2; GLAMOS, 2016; Lambiel and Talon, in press). In addition, between 1964 and 2009, where the glacier has not the completely disappeared from the study area, the glacier surface has lowered by c. -110 m (Figure 1b).

Today, the Otemma proglacial area is characterized by an alluvial plain reworked almost continually by a braided river and unvegetated hillslopes and Little Ice Age lateral moraines, about 200 m above the valley floor (Figure 1A). The moraines are, as is often the case in Alpine deglaciating valleys, a consequence of glacier advance in periods within the Little Ice Age (LIA) between 1350 and 1850. These moraines and the hillslopes beneath them have become progressively de-buttressed following glacier recession. The study is focused on a 300 m length of the right lateral moraine exposed South-South East (Figures 1A-square a, and 1B). Directly above the study area there is a moraine bastion, built up by the smaller Glacier de la Grande Lire (Figure 1A-square c) which accumulates sediments resulting from physical and chemical weathering, as well as the mechanical glacial erosion of granitic bedrock (gneiss with highly-foliated minerals such as quartzite and mica). Grain sizes range from sands to small boulders (Ø<2 m). This moraine bastion is clearly the main supply to the sediment cascade of the study area.

The moraine itself is partly discontinuous as there are some zones where moraine accumulation did not occur and there are bedrock exposures. Directly above the LIA limit, gravitational and torrential processes have reworked some sediments from the moraine bastion generating several debris accumulations (Figure 1A-square b). The work has a particular focus on two sediment sub-systems (Figure 1B). The first (SS1), located between two massive rock outcrops in the most down-valley part of the study area, at elevations between 2450 m a.s.l. and 2680 m a.s.l. (c. 41,700 m^2), is characterized by a well-developed torrential transport system connecting the moraine bastion to the alluvial cone (black dotted line in Figure 1B). In 2017, when the work was conducted, it was situated c. 870 m from the glacier terminus and it had a mean slope gradient of c. 62°. The second (SS2), situated c. 120 m up-valley from SS1, consists of a complex geomorphological system. It is located at elevations ranging from 2460 m a.s.l. to 2720 m a.s.l. (c. 39,500 m^2) and its sediment cascade is initiated by sediment accumulation at the head of a gully system (gray dotted line in Figure 1B). It is 750 m from the glacier terminus (based on the glacier's 2017 position) with a mean slope gradient of c. 67°, slightly greater than SS1.

2.2 Methodological approach

Evolution of the moraine was studied both quantitatively and qualitatively using historical aerial digital photogrammetry (Figure 3). Aerial imagery of the study area is available from the 1960s. So, we decided to focus on the decadal timescale, following Micheletti et al., (2015a) and Staines et al. (2015). To determine the geomorphic evolution of the hillslopes, archival digital photogrammetry was applied to historical aerial images to produce Digital Elevation Models (DEMs) for seven years (T_X). The associated DEMs can be affected by random and systematic errors, related to image quality, ground control point survey quality, surface composition, topographic complexity and interpolation methods (Lane et al., 1994; Lane, 1998; Hancock, 2006; Heritage et al., 2009), and this requires error management which is explained below. Once the DEMs were corrected, they were used for two different purposes. First, DEMs of Difference (DoDs) were used to quantify the patterns of erosion and deposition for different periods. Second, assuming hydrological control as an important factor in sediment reworking, we conducted a hydrologically-based sediment connectivity analysis (after Lane et al., 2017). The Lane et al. (2017) approach differs from the approach of Cavalli et al. (2013), because it seeks to make a distinction between DEM sinks that are likely to be due to noise in the DEMs and those that are likely to be real, a product of the structural organization of the landscape. These analyses were combined with detailed geomorphological information, conducted through geomorphological mapping and analysis of elevation profiles, with the aim of determining the evolution of the sediment cascade through time. All data generated in this project are available at ebibalpin.unil.ch.

2.3 Methods

2.3.1 Archival digital aerial photogrammetry

Photogrammetry is a long-established remote sensing technique that has now been proven for the reconstruction of long-term geomorphic changes in Alpine environments all over the world, including river-floodplain systems (e.g. Lane et al., 2010; Watanabe and Kawahara, 2016; Bakker and Lane, 2017), glaciers (e.g. Immerzeel et al., 2014; Gabbud et al., 2016; Mölg and Bolch, 2017), proglacial areas (e.g. Staines et al., 2005; Schiefer and Gilbert, 2007; Carrivick and Rushmer, 2009) and periglacial zones (e.g. Kääb and Vollmer, 2000; Micheletti et al., 2015a). The principles of applying photogrammetry to historical archival imagery are outlined in Lane et al. (1993). During the 1990s and 2000s, such approaches were developed to make use of digital datasets which meant that access to expensive hardware removed one of the major limits to the application of archival photogrammetry (Micheletti et al., 2015a). These limits have been further reduced by the development of SfM-MVS software which is both cheaper and faster than traditional digital photogrammetry software. However, applying SfM-MVS software to archival imagery requires a very different workflow to that associated with conventional SfM-MVS applications (e.g. Westoby et al., 2012; Fonstad et al., 2013) which use specially-acquired imagery. Bakker and Lane (2017) showed that because the number of images, and hence the extent of image overlap in archival datasets, is substantially lower than is normal with specially-acquired imagery, SfM-MVS software struggles to reconstruct the correct interior geometry of the cameras used for data acquisition. As a result, we followed more classical archival methods but applied them using SfM-MVS software.

The historical images used in this work were provided by the Swiss Federal Office of Topography (SwissTopo) for seven distinct dates between 1964 and 2009 (Table 1). Camera certificates were available for each period and, rather than determining them using the SfM-MVS software, we followed Bakker and Lane (2017), and fixed the calibrated focal lengths, and tangential and radial lens distortions as proscribed in the camera calibration certificates. The SfM-MVS software was then used to reconstruct the position and orientation of the cameras at the time of acquisition (Figure 4).

We use the commercial software Pix4D© for the photogrammetric analyses. The camera calibration certificate was used to specify the calibrated focal length and the tangential and radial lens distortion for each image. Working in the Swiss CH1903 coordinate system, we manually inserted the initial coordinates of the principal point position of every uploaded image provided by SwissTopo. To aid the initial solution of the collinearity equations, we included Ground Control Points (GCPs) (Fonstad et al., 2013). GCP collection followed Micheletti et al. (2015b). A Trimble© R10 dGPS was used establishing the base station on a fixed point and left recording at 1 minute intervals for 12 hours (Figure 5). These coordinates were then corrected *post hoc* using the continuously recording dGPS network SwissPos©, with data taken from the Lausanne and Zermatt stations. The base station precision was better than ±0.01 m in the horizontal and ±0.02 m in the vertical after correction. The stable points, commonly very large boulders and bedrock outcrops, were measured in a real-time kinematic survey, with data point precision of ±0.01 m in the horizontal and ±0.02 m in the vertical. These were also corrected *post hoc* to the SwissPos-corrected base station. The choice of GCP locations had the initial goals of: (i) a good distribution across the study area (Küng et al., 2011; Javernick et al., 2014) and (ii) long-term stability and accessibility (Micheletti et al., 2015b). The latter was the principal requirement; this meant in practice that we had to relax the former, but noting that with the SfM-MVS photogrammetric software, the bundle adjustment uses conjugate features identified from across the area of interest, so increasing the robustness of estimations in zones beyond where the GCPs were measured. In total, we were able to measure 13 stable GCPs. Due to the stability criterion, they are concentrated in the southern part of the LIA proglacial margin (Figure 5).

An initial block bundle adjustment solution was obtained using 3 GCPs. Then, additional GCPs were added manually to improve the solution, minimizing the Root Mean Square Error (RMSE), and a point cloud was generated. The bundle adjustment used between 4 and 8 GCPs (Table 2), which is a relatively small number compared to the total surface covered by the aerial images (c. 29,000 m^2), and a total number of tie-points per image between 36,341 and 93,388 points. If the bundle adjustment has been a success, the theoretical precision should be commensurate with or better than the mean RMSE, which it is for 1970 1977, 1983 and 2009. It is slightly degraded for 1964 and 1995 and more so for 1988. Some degradation was expected for 1964, as the number of tie-points used is relatively low, related to the poorer scan resolution (Table 2), which reduced image texture and also resulted in poorer DEM resolution and higher georeferencing error. The horizontal errors are generally low, but this is less the case for 1988, which explains the higher RMSE. It is likely that there is some error in precise positioning of GCPs during digitization which may also explain the higher σ_z (Table 2). Generally, higher resolution imagery with a better theoretical precision, produces lower σ_z, the exception being 1988 with degraded σ_z and 2009 with better than expected σ_z.

Even with acceptable RMSE values, with SfM-MVS software it is possible that the DEMs contain residual systematic error, such as tilt or doming (Bakker and Lane, 2017). Lane et al. (2004) showed that although this error is systematic, it commonly results from residual random uncertainty associated with parameters in the collinearity equations. Commonly, it is only notable when consecutive DEMs are compared (Lane et al., 2004). As a result, we randomly sampled 98 points from stable zones, the latter restricted to bedrock outcrops (Figure 6) and so should have constant z values. We considered the three dates that had the most GCPs and on this basis noted that 1983 had the best RMSE. Thus, 1983 was taken as the baseline. We then took the planform position of each of the 98 points and estimated the corresponding z values in the derived DEMs using the respective final bundle adjustment for each DEM. We compared all z elevations to the 1983 DEM z elevations. Table 3 (before correction) reveals that the mean error was high suggesting the presence of systematic error which would also inflate the standard deviations of error. This pointed to the need to remove systematic error before the DEMs could be used to calculate DEMs of difference and confirms the findings of Bakker and Lane (2007) in a similar archival application using SfM-MVS software.

2.3.2 Post-processing of point clouds

Systematic errors were removed by co-registration onto a single reference point cloud using CloudCompare© (Miller et al. 2008; Micheletti et al., 2015c; Bakker and Lane, 2017). To better manage the registration, we limited the point clouds to our study area with a total coverage of ca. 0.20 km^2 (Figure 1B). Visually, and also taking the arguments above, the best point cloud appeared to be that of 1983 and all point clouds were co-registered onto this one. We identified features that had to be stable during the study period (primarily bedrock outcrops but also large boulders) and used it in the iterative adjustment method in CloudCompare to minimize the RMSE between each point cloud and that for 1983 (table 4).

Final RMSEs values were comprised between 1 and 0.1 with a decreasing order from the oldest point cloud to the most recent (Table 4). Note that, even if the point cloud of 1964 had a higher point cloud density compared to others, notably 1988 and 1995, the resulting RMSE is the highest; it reflects the poor imagery resolution at the lower altitudes that did not allowed clear definition of geomorphic features.

After co-registration, 1 m resolution DEM grids were created in Arcmap 10 using kriging. This interpolator is recognized to be the best solution for complex landscape surface data because of its ability to conserve larger-scale topographic detail (Moore et al., 1991; Holmes et al., 2000). We chose to use an ordinary spherical semi-variogram leaving the software to find the best values for the fitting parameters (sill and nugget).

Finally, to reduce linear systematic DEM error further, we adopted a multi-regression method using stable zone elevation values in order to better fit individual DEMs to the reference. In practice, we compared observed elevations (Z^0) with those predicted by the mathematical model (Z^p) and, in order to increase their correlation coefficient (r) until a minimum reference value of 0.9, we iteratively deleted outlier values. Once calibrated the regression model has been applied to every cell composing the DEM (Table 5). The combination of point cloud registration and tilt removal resulted in a substantial decrease in systematic error and precision (Table 3)

Resulting DEMs were used to detect both historical surface geomorphological changes, to relate these to a geomorphological map of the area (Lambiel et al., 2016), and to quantify changes in hydrologically-driven sediment connectivity through time. Our focus is upon evolution within the Little Ice Age limit which we base upon a marked difference in vegetation development as well as the distribution of moraine ridges. Due to slope pressure releases triggered by deglaciation, and consequent mass movements, it is possible that ice surface altitude at the LIA maximum has been underestimated (Cossart et al., 2008). In the text, when we refer to the LIA limit, we assume that this is not the case.

2.3.3 Calculation of DEMs of difference

DEMs of difference (DoD) were calculated to determine erosion and deposition patterns (Lane et al., 2003). This latter is based on a simple matrix subtraction (T_{X2}-T_{X1}) between DEMs (Brasington et al., 2003; Lane et al., 2003). In practice, we then needed to account for residual uncertainties in the dataset (Brasington et al., 2003; Lane et al., 2003), related to (i) the quality of individual points in point clouds, (ii) the density of points available to represent the surface, (iii) the distribution of points within the study area and (iv) the interpolation method used to generate regular surface within points (Lane et al., 1994; Lane, 1998; Hancock, 2006; Heritage et al., 2009). In our case, we generated very high point densities, with very few zones of interest not covered, and so the primary focus was upon the uncertainty due to the quality of individual points. Thus, we used a simple error propagation method (after Lane et al., 2003) to investigate the propagation of error in individual DEMs of difference, which was defined as the level of detection needed for an elevation change to be judged as significant. After Lane et al. (2003) this was defined as:

$$\text{LoD} = \pm t\sqrt{(\sigma_{DEM1})^2 + (\sigma_{DEM2})^2} \tag{1}$$

where: LoD = Limit of Detection threshold
t = Student's confidence interval threshold value (1.96 at 95% or 1 at 68%)
σ_{DEM1} = standard deviation of DEM_1 Z error
σ_{DEM2} = standard deviation of DEM_2 Z error

The standard deviation of individual DEM errors was calculated from a sample of 98 manually identified Z values representing spatially distributed stable zones recognizable on all georeferenced DEMs (Figure 6). We took a confidence interval for *t* of 68% as higher confidence intervals tend to overlook smaller magnitude but spatially homogeneous changes (Wheaton et al., 2010) and because this threshold produced visually plausible patterns of erosion and deposition. The resulting LoD values (Table 6) were used to threshold the historical DoDs in order to show only statistically significant geomorphological changes; values within the LoDs are classified as no significant change between dates.

2.3.4 Hydrological sediment connectivity analysis

Sediment connectivity is defined as the degree to which sediments can flux through the landscape, and in particular between sediment sources and downstream areas (Cavalli et al., 2013). The focus in this manuscript is on hydrological sediment connectivity, that is the water-driven transfer of sediment between two different compartments of a catchment sediment cascade (Fryirs, 2013). We assume that a sediment disconnection follows from when a sediment sink forms and removes sediment from the cascade for varying lengths of time (Fryirs, 2013; Bracken et al., 2015). In our study, the connectivity analysis has two main related objectives. The first is to detect the evolution of sediment sources and sinks on hillslopes within the LIA limits, as well the water-related sediment transport paths, starting from the analysis of DoDs and orthoimages. DoDs are used to investigate the functional connectivity of the system (i.e. sediment transfer between compartments) and, through their comparison with orthoimages, quantify the main sediment pathways and their evolution through time. The second is more focused on the analysis of the structural (dis-)connectivity. Following Lane et al. (2017), it investigates the distinction between natural and artificial sediment disconnections and their evolution through time. Sediment disconnection may be confused with DEM-related noise (Borselli et al., 2008; Cavalli et al., 2013; Lane et al., 2017), especially in DEMs derived from archival imagery with reduced precision as compared with specially-collected imagery. Here, we define natural disconnection as the flow path obstruction by a geomorphological feature that leads to a reverse slope (e.g. due to glacial or paraglacial landforms), while artificial disconnection is that caused by DEM noise. The identification of sediment flow paths starting from DEM data is commonly undertaken by forcing flow accumulation through filling all depressions regardless of their nature, and which risks removing disconnection that is natural rather than artificial (Arnold, 2010). As we are interested in determining the impact of the evolution of the moraine morphology on sediment connectivity between 1964 and 2009, we applied the methodological approach proposed by Lane et al. (2017) which captures how the level of hydrological connectivity evolves as a DEM is progressively filled. If there is a transition from small to large upslope contributing areas at fills below or close to the DEM noise, it is likely that a flow path is hydrologically connected. Where the transition occurs at higher magnitudes of fill, so it becomes more likely that the disconnection is real, the greater the level of fill, the greater the probability.

Our approach uses the TopoToolbox (Schwanghart and Kuhn, 2010) to apply the Holmgren (1994) flow routing to calculate the upslope contributing areas (*A*: defined as the area upslope of a surface element that drains to that element; Rieger, 1998) with different levels of DEM fill. The Holmgren routing uses:

$$FD_{(j)} = \frac{\left(tanB_j\right)^x}{\sum_{j=1}^{n}\left(tanB_j\right)^x} \qquad (2)$$

where: $FD_{(j)}$ = proportion of hydrological flow in direction j
B_j = slope gradient between the central cell and the cell in direction j; with B_j set to zero if B_j is negative in the downslope direction
n = number of slopes positive in the downslope direction
x = variable exponent varying between 0 and infinity

Equation 2 determines for each DEM cell the proportion of flow going in a specific direction. For x=0 flow is equally distributed in all directions regardless of slope; for x=1 flow is proportionally divided to all downhill cells as a function of slope gradient (multi-directional flow) and for $x = \infty$ all runoff is directed along the line of steepest descent (i.e. D8 algorithm) (Quinn et al., 1991; Holmgren, 1994; Lane et al., 2017). In our study, we decided to use variable x values (x=1,2,4,8,16 and 32) to determine their effects. The fill value begins at 0.1

m and is then increased dyadically to a maximum of 102.4 m (i.e. we double the fill level starting from 0.1 m and ending with 102.4 m). In the case of artificial disconnection (dotted line in Figure 7), we would expect a rapid increase in *A* when the level of fill (*K*) reaches values comparable to DEM noise (σ_{2009}). However, in the presence of natural disconnection (solid line in Figure 7), the *A* increase may be later (*K*>DEM precision) and also achieved in a number of steps if there a number of natural disconnections of different magnitude. To apply the method, accumulation is determined for identified cross-sections on the hillslope. By comparing curves, it is possible to determine how connection evolves through time. If, at the most recent date, perfect connection (i.e. a rapid increase in *A to A1*) is reached at lower levels of *K*, it implies that hydrological connection has improved upstream through, for example, gullying ("within-basin connection" dotted line Figure 8). If, at the same time, the asymptote is reached at a higher value of *A,* i.e. *A2,* then it means that the geomorphic processes have been sufficient to improve connections upstream of the area associated with A1 ("basin extension"; dash-dotted line Figure 8). It is also possible that blockages along a flow path develop as a result of erosion or that geomorphic processes elsewhere capture some of the upslope contributing area of the basin, causing the value of *A* linked to the asymptote to decline. Likewise, the establishment of a new natural disconnection, for example the interruption of gully channels though a landslide dam, could cause the curve to shift towards a higher level of fill.

In these analyses, our focus is on two sets of data. The DEMs for 1964 and 2009 extended right up to the basin divides above the two focus study areas. For these we visualize the connectivity and quantify how it is changing with the connectivity analysis. For the other dates, only part of the possible upstream contributing areas was in the DEMs and so we do not include these, except for visualization. We make one exception to the latter. We quantify the distribution of upslope areas for all cells within each system. To make these comparable between dates, we delimit the cells using a mask defined by the smallest DEM extent in the time series, such that the areas being compared between dates are the same.

3. Results

3.1 Geomorphic change between 1964 and 2009

The visual evolution of the lateral moraine is showed in Figures 9 for 1964, the beginning of the study, and in Figure 11 for 2009, the end of the study. Figure 10 compares the elevation profiles of the subsystems. The LIA limit is shown as a discontinuous moraine ridge (Figure 9B). Above it, the hillslope is almost totally covered by a stable soil-vegetation complex, indicating relatively low levels of geomorphic activity, except for the Glacier de la Grande Lire's moraine bastion and its depositional zone immediately below. This zone is comprised of non-cohesive sediments, fines to small boulders (Ø<2 m), and it appears to be dynamic given the absence of vegetation (I in Figure 9A).

The spatial configuration of the hillslope in 1964 is an important aspect to consider because it allows the identification of potential sediment transport paths and buffers. The remains of the moraine crest, mainly visible on the rock outcrops, are likely to represent barriers for sediment connectivity in both subsystems (Figure 9B). Nevertheless, the geomorphological map (Figure 9B), the elevation profiles (Figure 9C and Figure 10) and the flow accumulation algorithm (Figure 9E) show that other potential obstructions to downward sediment transport are likely to be present. In SS1, we identified three more buffers (II to IV on Figures 9B, 9C and 10): the flatter zones located on the moraine bastion (II), the flatter area behind the LIA limit (III) and the concave fluvial scar just above the gully system (IV). In contrast, SS2 has a more complex profile resulting in the presence of more potential buffers (V to VIII on Figures 9B, 9C and 10) identified in: the flatter regions on the moraine bastion (V), the fluvial scar with multiple terraces behind the LIA limit (VI), the accumulation at the head of the gullies (VII) and the top of the supraglacial alluvial fan that would have formed on the glacier before its retreat (VIII). In general, all the above-mentioned obstacles are flatter zones constructed by the interaction between the bedrock outcrops and the reworked moraine sediment composing the debris-mantled slope during the LIA and post-LIA periods.

The situation in 2009 shows landscape evolution after glacier retreat and debuttressing (Figure 11). The most important geomorphological change concerns the growth of massive alluvial fans at the intersection between the hillslope and the alluvial plain once occupied by the Glacier d'Otemma (IX on Figure 11B). These fans may be related to greater upslope coupling, following from gully headward erosion activated in response to the glacier thinning, which progressively eroded the debris-mantled slope and accumulations previously identified as potential buffers (Figures 10 and 11C). This process led to the development of a torrential stream system able to couple even more distant sediment sources (Figure 11E). However, subsystems were affected by different intensities of gullying. In SS1 gullies developed upslope reaching the moraine bastion, while in SS2 they eroded without any important upward extension (Figure 11B).

Figure 12 quantifies these changes in terms of DEMs of difference for SS1 between 1964 and 2009. During the phase of glacier recession between 1960 and the mid-1970s, there is substantial glacier loss and thinning, of about 50 m (figure 1B), and hence debuttressing. This is likely to have resulted in both gully incision and headward erosion (I on Figures 12A and 12B). If this delivered more sediment to the alluvial fans below, it would explain their increasing basal area and total volume (II on Figures 12A and 12B). At the end of the 1970s, the glacier completely disappeared leading to exposure of the valley bottom (III on Figure 12C). But, the stabilization of the hydrological base level does not interrupt the geomorphic response to glacial debuttressing. Since the 1980s, there has been a more intense erosion reworking of c. -2 and -7 m of moraine material per decade, and the gully head extends into the moraine bastion and there was the removal of the intermediate sediment sink behind the LIA ridge (I on Figures 12D, 12E and 12F). At the same time, the alluvial fan experienced continuous aggradation (c. +2 and +7 m per decade; II on Figure 12D, 12E and 12F) only interrupted during the 1983-1988 period, when we recorded surface lowering rates between c. -2 and -9 m (IV on Figure 12D). Pioneer vegetation developed on the glacio-fluvial plain but not on the gully-alluvial fan system indicating that sediment reworking from the moraine bastion is still today relatively active.

Figure 13 shows the historical geomorphological evolution of SS2. In this case, the fluvial processes occurring in response to the glacial debuttressing are mainly located below the LIA limit, only partially affecting the flatter zone located above the gully system. During deglaciation, which for this part of the hillslope lasted until the end of the 1980s, Figure 13 suggests an initially weak fluvial activity (I on Figure 13A) followed by increased activity and alluvial fan erosion (I and II in Figure 13B and Figure 13C). This latter is concurrent with significant glacier thinning of c. 60 m, which occurred between the 1970s and the beginning of the 1980s (figure 1B). Note that until the exposure of the valley floor, even in this case, the alluvial fan shows erosional patterns because its aggradation rate was lower than the glacial thinning. From the mid-1980s, similar to SS1, base level fell to the valley floor altitude (III on Figure 13D) and headward gully erosion continued, with c. 2 to 6 m of reworked sediments deposited on the alluvial fan (I and II on Figure 13D, Figure 13E and Figure 13F). During the 1983-1988 period, up-valley, where the glacier thinning exposed a massive bedrock patch at the interaction with the alluvial plain (IV on Figure 13D), fluvially-driven erosion appeared to be less intense. The alluvial cone began to subdivide into two distinct fans. In the 1988-2009 period further fluvial erosion, encouraged by the complete disappearance of the glacier (III on Figure 13E), allowed sediment remobilization. Gully incision (I on Figures 13E and 13F) took place with two different spatial intensities: on the upslope section fluvial erosion reworked c. 6 m of moraine material exposing the underlying bedrock (IV on Figure 13E), while on the downslope section, even if recorded rates are higher (c. -15 m), complete denudation did not take place. The was major alluvial fan aggradation, as with SS1, but also some basal erosion (II on Figure 13E and V on Figure 13F) and this may have reactivated headward gully erosion (c. 10 m) in the most downslope area, where the moraine layer is still thick (I on Figure 13F). The reworked sediments deposited (c. +6 m) immediately at the alluvial fan top prevented bedrock exposure (VI on Figure 13F). Vegetative colonization took place on the alluvial scar, on the bedrock outcrops and on the alluvial plain without affecting the gully channels and related alluvial fan.

If we combine these patterns together, the total volume of remobilized sediments in the two studied systems was 1.8 km^3, with 1.1 km^3 related to SS2 and 0.68 km^3 to SS1. If we standardize these by their surface extents, both subsystem experienced intense erosion rates until 1988, between -10 mm/year and -716 mm/year for SS1; and -31 mm/year and -1002 mm/year for SS2. This period was followed by increased deposition with rates ranging from 21 to 298 mm/year for SS1 and 21 to 170 mm/year for SS2. In general, there is a clear trend towards decreasing sediment reworking rates over time (Figure 14).

3.2 Hydrologically-defined sediment connectivity

Given the geomorphic evolution described above, this section seeks to quantify the evolutoin of hydrological connectivity. Figure 15 shows the upslope contributing area for both the 1964 and the 2009 DEM calculated with all pits filled using a fill threshold defined by the σ_{2009} (Table 3). Figure 15 confirms that in 1964 there were drainage patterns present that can also be found in 2009, such as below the moraine bastion associated with the Glacier de la Grande Lire. The LIA moraine of the Glacier d'Otemma is not-well defined and continuous in 1964, such that there is no clear glacially inherited obstacle to hydrological connection. If there was one, it had been breached before 1960. In 2009, after complete glacial debuttressing, the efficiency of the hydrological sediment connectivity has generally increased as the hillslope is characterized by higher upslope contributing area values (Figures 15 and 16).

Distributions of upslope contributing area (Figure 16) confirm the visual assessment above. In SS1, cells having values between c. 0 and 10 m^2 (26.6% in 1964 and 18.5% in 2009) are clearly diminished over time in favor of cells with higher values, especially ranging from 10^4 to 10^6 m^2 which increased from 31.3% in 1964 to 37.8% in 2009). Mean cell values increased from 2750 m^2 in 1964 to 4790 m^2 in 2009 (+73.8 %). Histograms characterizing SS2 suggest trends similar to SS1 but with different intensities. Cells showing cell values of c. 0 and 10 m^2 (27% in 1964 and 18.8% in 2009) diminished markedly, while pixels values comprised between ca. 10^4 and 10^7 m^2 increased (38.9% in 1964 to 48.6% cells in 2009). In this subsystem, the mean cell values in 1964 was 2820 m^2 compared to 8320 m^2 in 2009 (+195 %). In general, cells moved towards an overall higher upslope contributing area value.

Figure 17, shows changes in connectivity between 1964 and 2009 and highlights that the increases in upslope contributing area appear to be related to the development of the gully network, encouraged by headward erosion in response to gradual glacier thinning (Figure 10), connecting previously-separated geomorphological compartments and eroding features once acting as barriers. In fact, the increase in connectivity is clearest along the gullies and on the alluvial plain. Results suggest that between 1964 and 2009, at the hillslope scale, headward erosion by gullies increased the total area of potential remobilizable sediments by c. 24,810 m^2 (c. +3%, Figure 17). Given this, in 2009, regions with small accumulation areas and disconnections, especially in regions located below the LIA limit, seem to be still significant (Figure 15). As a result of better connectivity, it is possible that increased flux of down-wasted sediments developed the alluvial fans at the hillslope base. Indeed, below the 1964 ice contact line, there is a tendency for flow to become more diffuse (Figure 15 and 17) suggesting a reduction in connectivity from this fan development.

Figure 16 also suggests that the evolution of connectivity was progressive and in some cases could be discontinuous with increases in disconnection following increases in connection. Histograms of upslope contributing area show that the total number of cells having small values, attributable to a less efficient connectivity, experience a strong increase during the 1977-1995 period. For SS1 there is an increase of the number of cells (+16%, equivalent in area to 1252.6 m^2) with values comprised between 10^2 and 10^3. On the other hand, this increase (9.4% equivalent to 1919.7 m^2) concerned cells having values between 100 and $1x10^4$ m^2. Figure 18 clearly shows this phenomenon highlighting that at the same time as the activation of new sediment sources through gullying, others regions, essentially located near the main channel have become less connected (Figure 18D, 18E and 18F). The connectivity clearly improved in 2009 when upslope contributing area values again evolved towards higher values (Figures 16 and 18G).

The evolution of connectivity between 1964 and 2009 is quantified in Figure 16 in terms of the relationship between the degree of pit filling and the upslope contributing area for different values of the Holmgren routing parameters (Equation 2). In 1964, the sediment connectivity in our study area was characterized by a general disconnectivity both above and below the LIA limit. Figure 19 shows that all sampled locations have one or more abrupt increases in basin contributing area reflecting the potential barriers (or sinks) identified in Figures 9, 10 and 15. Geomorphological analysis and the upslope contributing area map in 1964 showed that these locations were characterized by both complex topography (such as reverse slopes) and abrupt decreases in accumulated area immediately downstream . Given this, is it also important to note that all transitions are at levels of fill greater than the DEM related noise suggested by the σ_{2009} (Table 4), demonstrating the probable presence of natural rather than artificial disconnections.

The quantification of the buffers affecting the hydrological sediment connectivity of SS1 in 1964 (samples I-III on Figure 19) did not suggest significant sinks. At the LIA limit region (I on Figure 19) and on the moraine bastion (II in Figure 19), accumulated areas are c. 5.6 x 10^5 and 9.3 x 10^4 m^2 respectively and perfect connection is achieved through fills of 0.8 m and 6.4 m, respectively. In contrast, the hydrological connectivity on the alluvial fan (III in

Figure 19) is characterized by two increases at 3.2 m and 12.8 m starting from an accumulated area value of 7.3 x 10^4 m^2, suggesting the presence of more complex topography. In 2009, the hydrological sediment connectivity has increased as the fill levels necessary to overcome obstacles are no longer present. Values for the accumulated area remained stable at the LIA limit, suggesting that perfect connection was reached in 1964, while it slightly increased for the moraine bastion (7 x 10^3 m^2) and the alluvial fan (2.7 x 10^4 m^2) (Table 7).

On the other hand, the SS2 data in 1964 (IV-IX on Figure 19) are all characterized by two or more increases in upslope contributing area, between 3.2-6.4 m and 12.8-25.6 m of fill, except for the moraine bastion summit (VIII on Figure 19) which has an additional small increase at 0.8 m. This fact highlights that, at the beginning of the study, numerous buffers affected the fluvial system making SS2 more disconnected than SS1. Concerning the Holmgren parameter, in general we find a more concentrated flow on the moraine bastion compared to the lower LIA section. The subsystem in 1964 was characterized by a wide range of accumulated area comprise between 1 x 10^2 m^2 for the moraine bastion ridge and 6.3 x 10^5 m^2 for its central part (Figure 19 and Table 7). In 2009, the efficiency of connectivity increased and most of the sampled locations suggest the removal of barriers (IV, VI, VIII, IX on Figure 19) or a reduction in their importance (V and VII on Figure 19). In these latter cases, buffers are described by a fill level of 6.4 m and their continued existence is probably due to, respectively, bedrock exposure and the crest of the moraine bastion. Nevertheless, despite the limited extension of the gully shown on Figure 13, gully erosion may have been important, especially on the lower LIA section, while on the moraine bastion it is the upslope basin reorganization which drove the improvement of connectivity. The combination of these mechanisms allowed major changes in the spatial configuration of the geomorphic subsystem and the transition to a more diffuse flow. However, note that in the LIA limit region, flow is concentrated and it may be related to recent headward erosion that affects the flatter accumulation at the gully head (I on Figure 13F). In contrast, the markedly diffusive flow characterizing the alluvial fan (IX on Figure 19) confirms how its formation is likely to have reduce the connectivity between the slope and the proglacial margin. Changes in accumulated area supports the above observation. In summary, in all zones where gullying dominated (V, VI and VII in Figure 15), except for the LIA region were the situation remained stationary, there was an increase in accumulated area; whilst in zones of deposition (VIII and IX in Figure 15) reductions were noted (Table 7).

4. Discussion

4.1 Glacier dynamics and the geomorphic response of the lateral moraine

The Glacier d'Otemma has been in continuous retreat since the Little Ice Age (Figure 1 and 2), leading to the progressive debuttressing of its lateral moraines. Debuttressing involves both glacier retreat and progressive thinning of ice, changes that lead to a local hydrological base level fall and which may trigger paraglacial stress release and mass movement (Cossart and Fort, 2008; Porter et al., 2010). As the glacier terminus moves up valley, there was a tendency for up slope gully erosion of between -10 mm/year and -1002 mm/year and deposition of reworked sediments with rates of ca. +21 mm/year and +298 mm/year on the glacier surface, or valley bottom, in the form of alluvial fans. In some cases, especially in SS2, erosion to bedrock was observed (Figure 13E). Fluvial incision rates are greater than measurements of gully erosion on recently deglaciated terrains in Norway (30 mm/year and 170 mm/year; Ballantyne and Benn, 1994; Curry 1999) and the Swiss Alps (49 mm/year and 151 mm/year; Curry et al., 2006). Higher recorded rates in our study area may be related to a difference geomorphological context, especially the presence of a moraine bastion or a thicker till layer covering the slope left by the glacier, compared to previous works. It may also reflect the exceptionally high rates of glacier recession. In contrast, measured rates of alluvial fan aggradation fall within limits for Norwegian sites (8 mm/year and 44 mm/year; Ballanytne, 1995) and with measurements recorded in Patagonian sites (330 mm/year and 400 mm/year; Harrison and Winchester, 1997). With glacier recession, and associated ice surface thinning, the local base level falls, providing some temporary sediment storage for the sediment cascade along the valley flank (Cossart and Fort, 2008). After deglaciation, fans regulate the hydrological base level of the lateral moraine, but the response of the sidewall continued also once the glacier terminus had moved further upstream (Figures 12D, 12E, 12F and 13F). Erosion of alluvial fan toes was ultimately controlled by fluvial activity in the alluvial plain as observed more generally (Harvey, 1977, 1996; Lisenby and Fryirs, 2017). Such erosion could locally increase fan slope and potentially also encourage further gully erosion headward. Thus, slope response after deglaciation will not only be related to glacial debuttressing, causing glacial unloading and subsequent substrate cohesion loss, but wider landscape adjustment to non-glacial conditions, independent of paraglacial stress release (Cossart et al., 2008).

The length of time by which a glacial landscape responds after glacier recession has been described by Cruden and Hu (1993) using a theoretical exhaustion model. Starting from this mathematical simulation, Ballantyne (1995) and Curry (1999) argued that debris-mantled slopes complete their paraglacial cycle rapidly, in 50-200 years since deglaciation, and achieve a final, stable, geomorphic state characterized by an upper zone of exposed bedrock resulting from downslope sediment evacuation, a central zone where water flows in a concentrated way and several coalescent alluvial fans at the hillslope base (Chorley and Kennedy, 1971; Schrott et al., 2003). These studies were conducted for lateral moraines (e.g. Fåbergstølsdalen and Bas Glacier d’Arolla’s proglacial areas) that were not influenced by other effects, including the effects of the progressive reworking of moraine material upon underlying bedrock exposure. When perfect connectivity is reached at the geomorphic system scale and all sediment sources have been exhausted, the paraglacial period may be considered over. In such systems, sediment inputs may still occur but they become limited to rockfalls linked to paraglacial and paraperiglacial processes (Mercier, 2008). In our case, the two studied subsystems were located just below a moraine bastion that clearly acted as a primary sediment source. Due to its persistence, it is logical to expect that the time needed to achieve a perfect connection and to rework available unconsolidated sediments is longer. In 1964, the hillslope was characterized by multiple buffers (or sinks). In 2009, after the headward erosion gullies some of them were still present, especially on the moraine bastion (Figure 19), and sediment reworking towards the alluvial fans continued to occur (Figures 12F and 13F). Once all the remaining buffers are eroded, additional time will be needed to exhaust the newly activated sediment sources (Cossart, 2008). Thus, following Ballantyne (2002a) and Cossart (2008), we are unable to predict the duration of the paraglacial period in our study area using the equation formulated by Cruden and Hu (1993). This is because both subsystems are still perturbed by base level change (i.e. alluvial fan basal erosion) and headward erosion, which in combination unlock new sediment source areas and sediment transfers in the form of secondary peaks.

4.2 The sediment cascade and hydrological connectivity changes following glacial debuttressing

As identified in other cases (e.g. Cossart and Fort, 2008), at the beginning of our study the sediment cascades associated with the subsystems were affected by several intermediate sinks that obstructed sediment flux to the valley bottom (Figures 11) and diminished their effective catchment areas (Harvey, 2002). These latter were composed of inherited glacial landforms, including a discontinuous LIA ridge that partially breaks up the slope, as also observed by others (e.g. Knight et al. 2007; Cossart and Fort, 2008; Lane et al., 2017), and by barriers associated with interactions between paraglacial sediment reworking and local topography (Cossart, 2008). Sediment connectivity is dynamic because it changes in response to erosion and deposition, processes which itself regulates (Hugget, 2007; Fryirs, 2013). In our case, we have identified both connectivity-increasing and connectivity-decreasing processes, in the form of erosion of gullies and deposition of alluvial fans respectively (Figures 12, 13 and 19; Cossart and Fort, 2008). Progressive disappearance of intermediate sinks allowed an increase of hydrological connectivity, with mean upslope contributing area increasing in each subsystem by between +73.8% and +195.8%, respectively (Figure 16, 17 and 19). This process appears to be linked to base level fall following gradual glacier thinning (Figures 12 and 13) coupled with, as reported by Cossart and Fort (2008) in their work on the Vallouise valley, the accumulation of meltwater coming from a hanging glacier (de la Grande Lire, Figure 1) promoting pronounced gullying. Our results therefore confirm the conclusions of Lane et al. (2017) that gullying can be an important process in breaching features that disconnect downslope sediment flux, such as moraine ridges. Increases in connectivity have a second effect: not only do they allow flux more readily through a point in the landscape; they can also reorganize drainage basin areas upstream, potentially increasing (or decreasing) upslope contributing area and in turn increasing (or decreasing) erosion potential (Figure 19; Cossart, 2008; Fryirs, 2013; Bracken et al., 2015). In 2009, we reported that increases in accumulated area at sampled sites initially acting as buffers were between 2.1 x 10^3 m^2 and 7.0 x 10^3 m^2 (Figure 19 and table 8). At the LIA limit no increases occurred suggesting that in 1964 the LIA moraine had already been breached (Figure 19).

In parallel to the fact that gullying is the main driver that increases access to upslope sediment source via the erosion of buffers, our results also show that if headward erosion becomes too intense, the regions close to the gully channel can experience a reduction of their upslope contribution area and become less connected (Figure 18). The geomorphic evolution of the subsystems suggests that gully channels had high erosion, especially between 1970 and 1995, which established a well-defined gully network (Figures 12 and

13). The main channel was characterized by the greatest accumulated area values (> 10^9 m^2) meaning that the connectivity was at its maximum. During the same period, areas with lower upslope contributing area (10^2 to 10^4 m^2) clearly increased in area (by +16% for SS1 and +9.4% for SS2) suggesting that some regions have been decoupled and that their contribution to the downstream transport of sediment is strongly reduced (Figure 18). Following a geomorphological analysis, these are primarily gully sidewalls and patches of exposed bedrock, but it is also possible that gully capture elsewhere occurred to reduce the drainage through these sites (Figures 12 and 13). This may explain the fall in upslope contributing area at the moraine bastion ridge between 1964 and 2009 (XIII I Figure 19). Connectivity in these areas could be enhanced and re-established only if some depositional events were triggered, such as gully sidewall collapses or debris flow deposition, as occurred in 2009 leading to a reconfiguration of the gully network (Figures 12, 13 and 18; Curry et al., 2009; Eyles et al., 1988). This is a clear observation that gullying does not only lead to increases in sediment connection; erosion to bedrock may lead to the formation of sinks if the bedrock topography allows it, reducing connection. As Cossart (2008) observed, the evolution of connectivity on hillslopes can lead to morphological changes that may both maintain and reduce connectivity according to their setting with respect to sediment sources. Alluvial fans are the final accumulation zones where mass movements deposit reworked sediments because it is rare that debris flows triggered on sidewalls directly reach the alluvial plain (Becht, 1995; Curry, 2000; Hilger et al., 2019), even if the question is debated as sediment delivery to the main channel network in proglacial settings may be variable (Haas et al., 2012). Schrott et al. (2006) studied alluvial fan buffering along a valley, and found a clear relationship between deglaciation time and fan effectiveness in storing sediments. In the lower parts of deglaciated catchments, alluvial fans reduced hillslope to valley floor connection because stabilization also resulted in progressive revegetation. In contrast, in the upper parts of deglaciated catchments, even if alluvial fans intercept down-wasting sediments and reduce the overall connectivity, their sink effect is less pronounced due to the abundant amounts of unconsolidated glacigenic sediment prone to be reworked. The alluvial fans, for the most recent dates reported here, remain poorly vegetated suggesting that they may still be active. However, even without vegetation, alluvial fans have reduced slopes (Figure 11), tend to be diffusive in hydrological terms and are commonly well-drained (Figures 15 and 19), all of which will lead to their effectiveness in decoupling valley side slopes from the stream bottom (Fryirs and Brierley, 1999; Harvey, 2001; Lane et al., 2017). For instance, by 2009 no more obstacles influenced connectivity our subsystems but it is clear that the diffusive flow, mainly induced by gradient reduction, may encourage deposition before sediments reach the valley bottom (Figure 15 and III and IX in Figure 19). This observation concerns especially SS2, in which the gully is less channelized on the alluvial fan compared to SS1 (Gomez and Purdie, 2018). It may also explain the different behavior in terms of accumulated area evolution (deposition of 2.7 x 10^4 m^2 for sample III and erosion of 5 x 10^{-3} m^2 for sample IX, Figure 19 and table 7) between the two subsystems.

5. Conclusions

Observations of the historical geomorphological changes that occur during the glacial debuttressing phase of a retreating, temperate, Alpine valley glacier allowed us to quantify the evolution of two subsystems associated with valley side slopes. Both were below a Little Ice Age formed lateral moraine. We quantified this evolution in terms of historical surface change and hydrological sediment connectivity.

Due to gully development with erosion rates of -10 mm/year and -1002 mm/year, between 1964 and 2009 the two subsystems evolved into a sediment cascade resembling that of a torrent system: an upper source zone, related to a moraine bastion and moraine material covering the valley sidewall; a gullied channel mainly activated during high magnitude rainfall events; and, finally, a lower accumulation zone consisting of alluvial fans and the proglacial alluvial plain (with deposition rates of +21 mm/year and +298 mm/year). The evolution of these two subsystems then caused feedbacks that in turn impacted the evolution of their connectivity. SS1 evolved though the removal of intermediate sinks and an increase in connectivity (mean upslope contributing area value increased of 74%). Gully extension to the moraine bastion above maintained sediment supply and prevented erosion to bedrock and the development of bedrock-related steps that would have disconnected sediment flux. SS2, also experienced local increases in connectivity (increase of the mean accumulated area of 195%), but resulting increases in upstream sediment supply were insufficient to prevent denudation to bedrock, resulting in disconnection. In both cases, fluvially down-wasted sediments deposited at the hillslope base generated diffuse flow and disconnection, probably linked to the absence of a well-defined channelized stream.

We confirmed that gully headward erosion is the most important geomorphological agent to improve connectivity in this kind of environment, reaching even more upslope sediment sources and eroding buffers on its path (vertical connectivity). We quantified that, at the hillslope scale between 1964 and 2009, an area of 24810 m^2 (ca. +3%) had been "activated". However, we also highlighted that if incision is too intense, regions located close to the gully network (16% of the total area for SS1 and 9.4% for SS2) experience reductions in upslope contributing area (lateral connectivity). The main causes could be linked to gully sidewall formation, patches of bedrock exposure and gully capture events.

Thus, it is likely that the paraglacial phase contains subsystems that are responding to deglaciation at different rates, according to the extent to which erosion results in negative feedbacks that reduce sediment flux. This is a local effect, meaning that the prediction of the geomorphic response to debuttressing is context specific. We highlighted that in complex geomorphic systems, external sediment sources, in our case the moraine bastion, certainly perturb the evolution of the subsystems by promoting inputs of upslope originating sediment and forming secondary peaks (ex. following the activation of new sediment sources), prolonging the length of the paraglacial period. This is a hypothesis that merits further testing.

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Table 1: Main characteristics of the aerial imagery used and associated camera/image parameters.

Year	Camera type	# images	Overlap [%]	Scale [1:x]	Scan resolution [μm]	Ground Pixel resolution [m]	Calibrated focal length [mm]
1964	Lens 29 11.5 AG	14	60	1:17700	21	0.37	115.29
1970	Lens 174 15 AG	9	60	1:8600	21	0.18	152.10
1977	Lens 3004 15 UAG II	17	60	1:8600	14	0.12	153.02
1983	Lens 13018 15/4 UAG	17	60	1:8600	14	0.12	153.37
1988	Lens 13018 15/4 UAG	18	60	1:8600	14	0.12	153.37
1995	Lens 13220 15/4 UAG-S	13	80	1:8600	14	0.12	152.52
2009	Lens 13220 15/4 UAG-S	14	80	1:19800	14	0.19	152.52

Table 2: Pix4D's parameters and georeferencing error for stable zones marked with GCPs.

# GCPs	Tie-points per image	Resultant DEM resolution [m]	Theoretical precision [m]	Mean RMSE [m]	σ error [m]	
					X	Y
4	36341	1.998	±0.37	±0.457	±0.193	±0.245
5	81137	0.703	±0.18	±0.062	±0.01	±0.028
5	87125	0.608	±0.12	±0.104	±0.048	±0.190
8	92290	0.611	±0.12	±0.114	±0.085	±0.032
8	91906	0.631	±0.12	±0.369	±0.427	±0.093
8	93388	0.641	±0.12	±0.180	±0.156	±0.151
4	79913	1.651	±0.27	±0.081	±0.031	±0.023

Year	Error [m]			
	Before correction		After correction	
	$\bar{Z}$	σ (precision)	$\bar{Z}$	σ (precision)
1964	-18.8	±4.0	-0.9	±3.0
1970	-27.5	±4.9	-0.4	±0.5
1977	-9.9	±4.1	+0.4	±0.5
1983	-	-	-	-
1988	+4.9	±3.4	-0.1	±1.5
1995	+21.1	±6.8	-1.0	±1.3
2009	+1.3	±3.6	-0.6	±0.5

Table 3: Systematic error analysis based on stable zones.

Year	Number of points in point clouds	Point cloud densities [pts/m²]	Number of stable points used for registration	RMSE after registration [m]
1964	4'368’086	23.4	6	±1.00
1970	12'348’682	66.2	12	±0.90
1977	6'518’009	34.9	7	±0.58
1983	8'715’575	46.3	-	-
1988	2'858’505	15.3	9	±0.10
1995	3'045’655	16.3	8	±0.28
2009	7'524’428	40.3	8	±0.16

Table 4: Point cloud densities and co-registration characteristics.

Table 5: Correlation and regression coefficient of the multi-regression model.

Year	r	Regression equation ($Z^p=a+bZ^0+cZ^0$)		
		a	b	c
1964	0.9	36.5	-0.006	0.04
1970	0.92	-1.7×10^3	0.002	0.007
1977	0.94	-1.7×10^3	0.002	0.007
1983	-	-	-	-
1988	0.93	-1.1×10^3	0.005	-0.02
1995	0.9	-269.4	0.003	-0.02
2009	0.9	-2.2×10^3	0.004	-0.003

Year	68% LoD
1964-1970	±2.99
1970-1977	±1.70
1977-1983	±1.76

1983-1988	±1.74
1988-1995	±2.03
1995-2009	±1.78

Table 6: Limits of Detection (LoD) for the DEMs of Difference.

Location	Upslope contributing area (m²)		
	Date		
	1964	**2009**	**Change**
I	$5.6x10^5$	$5.6x10^5$	0
II	$9.3x10^4$	$1.0x10^5$	$+7.0x10^3$
III	$7.3x10^4$	$1.0x10^5$	$+2.7x10^4$
IV	$8.0x10^4$	$8.0x10^4$	0
V	$9.0x10^4$	$3.0x10^5$	$+2.1x10^5$
VI	$2.0x10^4$	$9.0x10^4$	$+7.0x10^4$
VII	$1.0x10^2$	$1.0x10^4$	$+9.9x10^3$
VIII	$6.3x10^5$	$1.4x10^5$	$-4.9x10^5$
IX	$2.0x10^4$	$1.5x10^4$	$-5.0x10^3$

able 7: Upslope contributing area values (1964, 2009 and difference between the two dates) for the sampled locations highlighted in figures 15, and 18.

Cell value classes							
10^5	10^5-10^6	10^6-10^7	10^7-10^8	10^8-10^9	10^9-10^{10}	10^{10}-10^{11}	10^{11}-10^{12}
61 9.7)	36224 (5795.8)	20603 (3296.5)	10093 (1614.9)	4891 (782.5)	1916 (306.6)	220 (35.2)	2 (0.32)
48 5.7)	38342 (6134.7)	20314 (3250.2)	8541 (1366.6)	4152 (664.3)	1935 (309.6)	283 (45.3)	5 (0.8)
89 1.2)	37709 (6033.4)	20137 (3221.9)	8516 (1362.6)	4259 (681.5)	1600 (256)	358 (57.3)	20 (3.2)
22 9.5)	31167 (4986.7)	17048 (2727.7)	6879 (1100.6)	3415 (546.4)	1610 (257.6)	459 (73.4)	13 (2.1)
66 4.6)	32836 (5253.8)	16635 (2661.6)	6816 (1090.6)	3273 (523.7)	1505 (240.8)	627 (100.3)	54 (8.64)

26 (2.1)	25610 (4097.6)	16263 (2602.1)	6247 (999.5)	2978 (476.5)	1559 (249.4)	569 (91.1)	98 (15.7)
48 (3.7)	41253 (6600.5)	16958 (2713.3)	6903 (1104.5)	4235 (677.6)	1968 (314.9)	336 (53.7)	1 (0.16)

ble A1. Difference in total number of cells per class and relative area (number of cells x [DEM resolution]^2, in brackets) for SS1.

ble A2. Difference in total number of cells per class and relative area (number of cells x [DEM resolution]^2, in brackets) for SS2.

	Cell value classes									
Year	**0-10^1**	**10^2-10^3**	**10^3-10^4**	**10^4-10^5**	**10^5-10^6**	**10^6-10^7**	**10^7-10^8**	**10^8-10^9**	**10^9-10^{10}**	**10^{10}-10^{11}**
1964	18524 (2963.8)	24057 (3849.1)	35065 (5610.4)	43490 (6958.4)	41229 (6596.6)	28637 (4581.9)	13862 (2217.9)	5506 (880.9)	1601 (256.2)	251 (40.2)
1970	14034 (2245.4)	22267 (3562.7)	35612 (5697.9)	45389 (7262.2)	15669 (2507.1)	32789 (5246.2)	15397 (2463.5)	4910 (785.6)	1318 (210.9)	72 (11.5)
1977	15071 (2411.4)	22952 (3672.3)	34870 (5579.2)	44592 (7134.7)	44713 (7154.1)	33239 (5318.2)	15539 (2486.2)	5256 (840.9)	1175 (188)	59 (9.5)
1983	26346 (4215.4)	31650 (5064)	39779 (6364.6)	42138 (6742.1)	35851 (5736.2)	23638 (3782.1)	11284 (1805.5)	4020 (643.2)	1183 (189.3)	87 (13.9)
1988	19988 (3198.1)	28343 (4534.9)	39634 (6341.4)	46827 (7492.3)	43638 (6982.1)	30105 (4816.8)	14727 (2356.3)	4793 (766.9)	1478 (236.5)	168 (26.9)
1995	19803 (3168.5)	29052 (4648.3)	40768 (6522.9)	48228 (7716.5)	43581 (6972.9)	30009 (4801.4)	13801 (2208.2)	5058 (809.3)	1513 (242.1)	126 (20.2)
2009	13165 (2106.4)	35995 (5759.2)	38235 (6117.6)	51752 (8280.3)	51627 (8260.3)	37870 (6059.2)	14778 (2364.5)	4303 (688.5)	1240 (198.4)	46 (7.4)

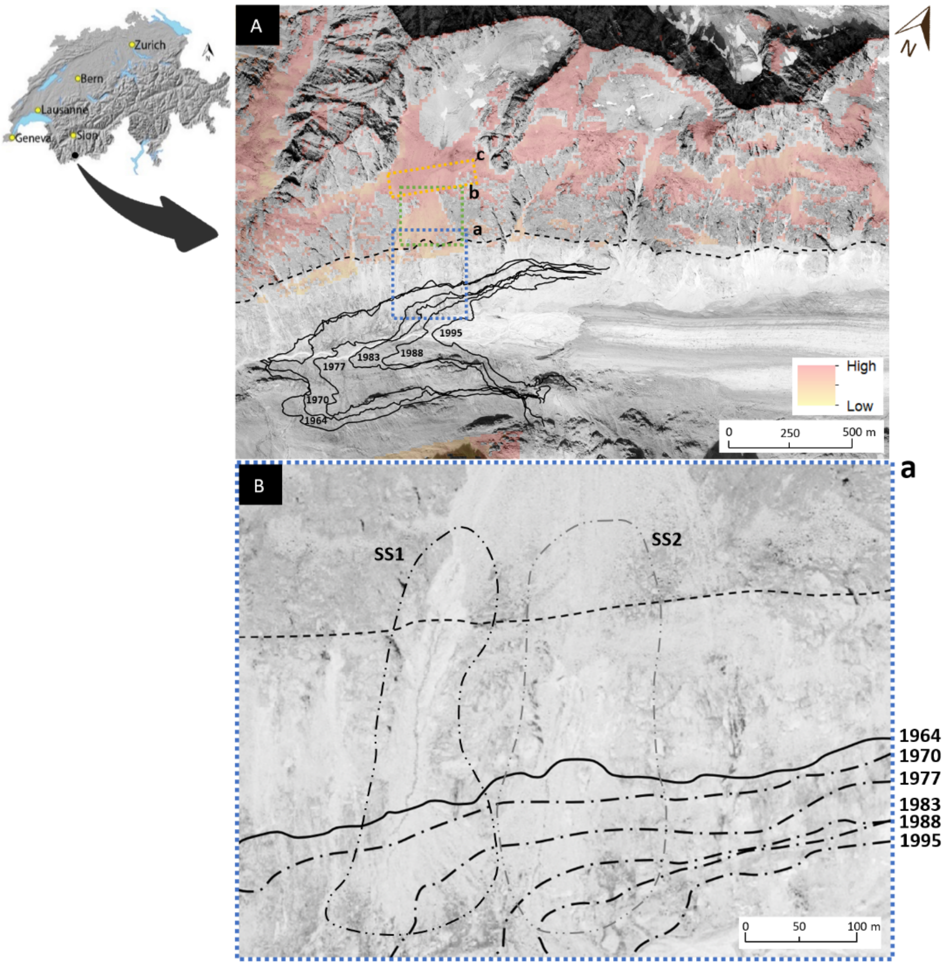

Figure 1: Location of the study area within the Glacier d'Otemma catchment (A) in 2009 and image of the valley sidewall in 2017 (B). Dashed line represents the supposed LIA right limit based on geomorphic investigation (location of moraine ridges and limit of vegetation). Squares in A highlight the LIA sector where our study is focused (a), the debris accumulation above the LIA limit (b), and the Glacier de la Grand Lire morainic bastion (c); solid lines refers to historical glacier outlines. Pixel colors refers to permafrost probability distribution (Deluigi et al., 2017). In B dashed solid and dash-dotted lines show the surface glacier thinning over time and the double dash-dotted lines represent SS1 and SS2.

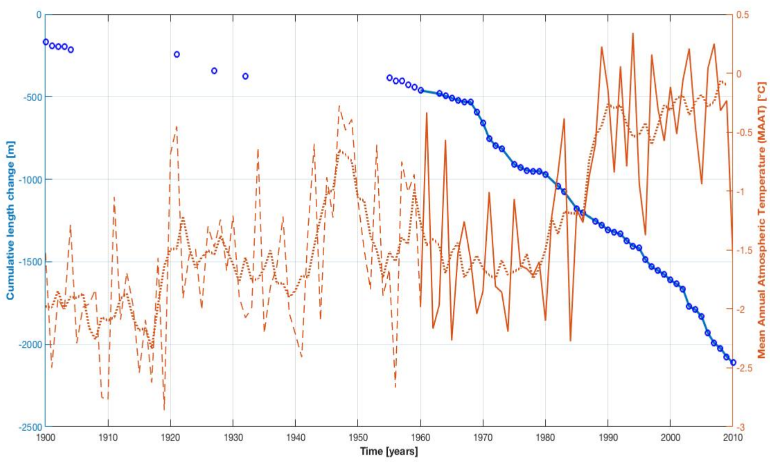

Figure 2: Glacier d'Otemma cumulative retreat in relation to the mean atmospheric annual temperature (MAAT) since 1900. Historical MAATs data refers to the Grand-Saint-Bernard climate station, located only 19 km from the study site at a similar altitude similar (2472 m a.s.l.) and all records are homogenized (the records are corrected by MétéoSuisse to remove influences of measurement method, small changes in station location, and changes in local climate influence such as urban development). The orange dotted line show the 5 years moving mean for MAAT.

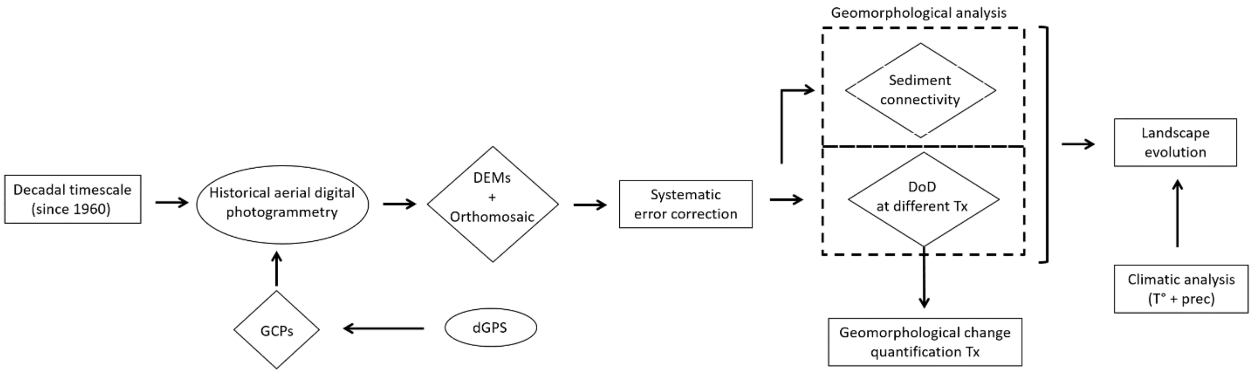

Figure 3: Schematic view of the methodology.

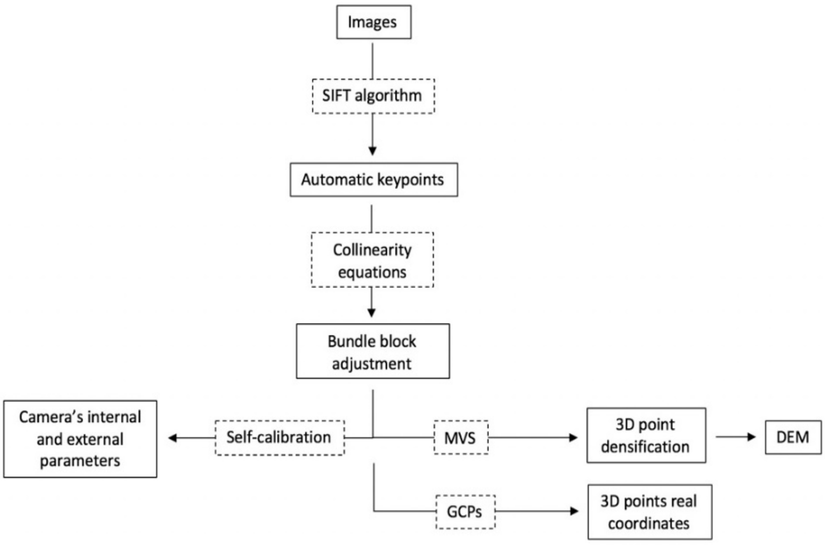

Figure 4: Schematic workflow for SfM photogrammetry, based on Fonstad et al., (2013).

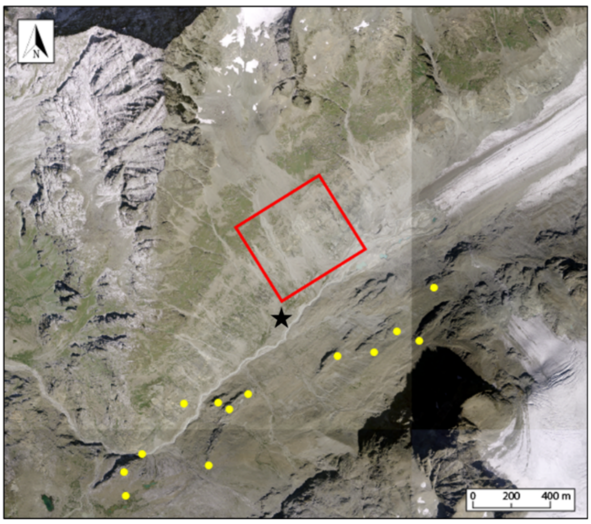

Figure 5: Spatial distribution of GCPs (points), base station position (star) and location of the study area (square). Coordinates of base station (in CH1903+) are 2'597'518.2 N, 1'086'726.2 E. Source of background image: SwissTopo.

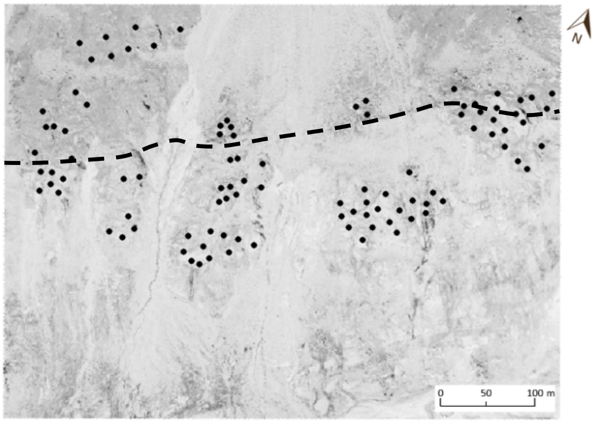

Figure 6: Location of the 98 Z points used to the detection of the systematic error and the computaion of LoDs (background image is dated 2009). Dashed line represents the supposed LIA limit based on geomorphic investigation (location of moraine ridges and limit of vegetation).

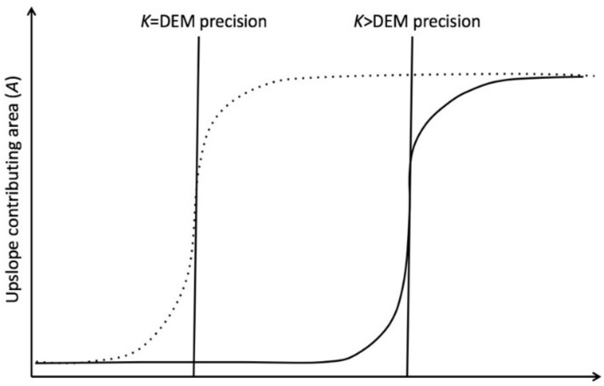

Figure 7: Hydrological sediment connectivity model for artificial (dotted line) and natural (dash-dotted line) disconnectivity in function of upslope contributing area (A), fill level and DEM precision (K) (Lane et al., 2018 – modified).

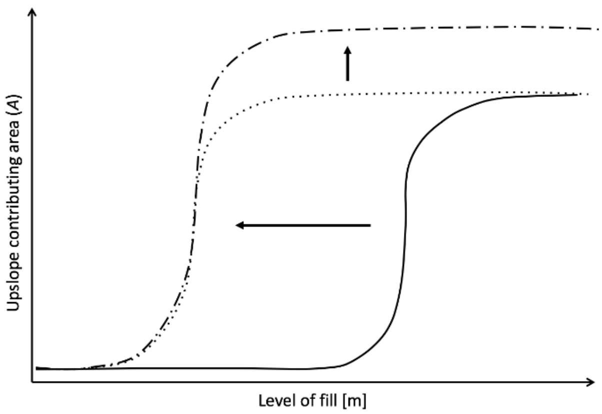

Figure 8: Hydrological sediment connectivity evolution ; base level fall and headward erosion propagation (dotted line) and upslope basin reorganization (dash-dotted line) (Lane et al., 2018 – modified).

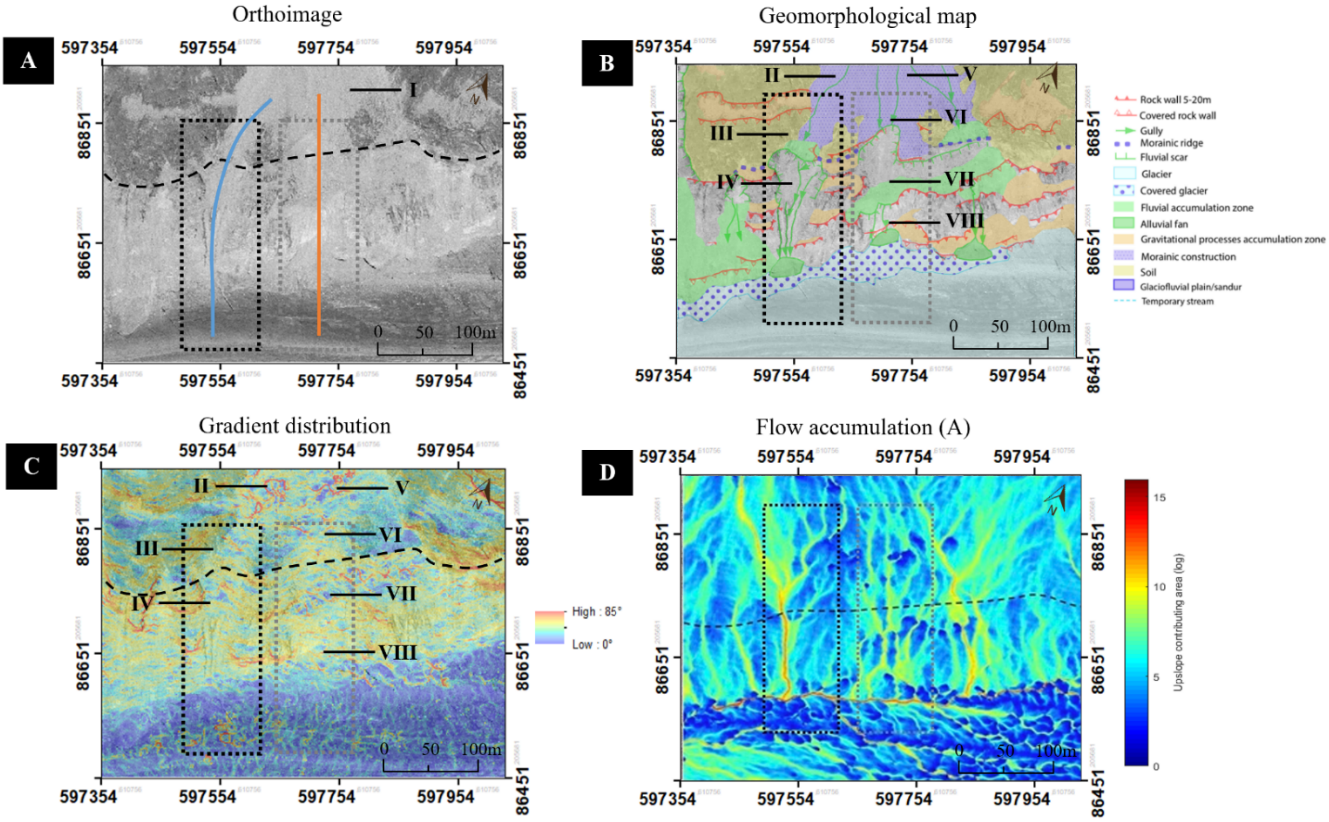

Figure 9: Geomorphic characteristics of the lateral moraine in 1964. A) Orthoimage highlighting the location of the elevation profile in Figure 10, B) geomorphologic map using the University of Lausanne's classification (Lambiel et al., 2016), C) slope gradient and potential buffers, D) flow accumulation algorithm calculated with all pits filled with a threshold defined by the σ_{2009} (Table 3). Dotted squares refer to SS location while dashed line represent the supposed LIA limit.

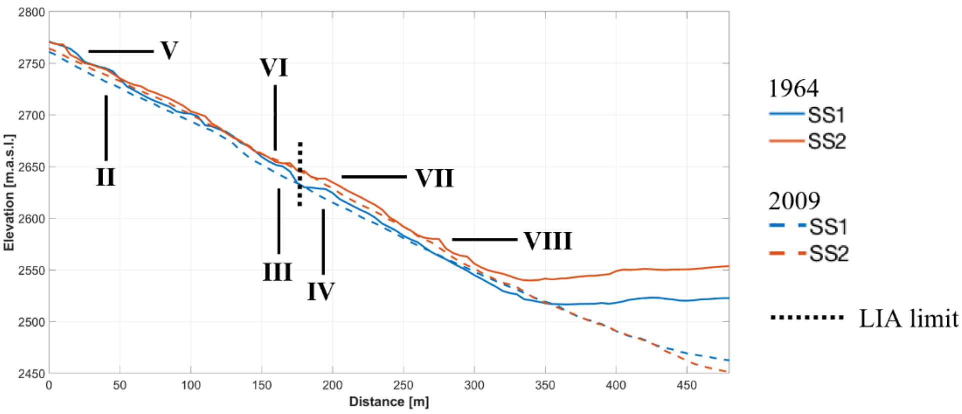

Figure 10: Gradient profiles of SS1 and SS2 in 1964 (solid lines) and 2009 (dashed lines). Spatial positions of the profiles are displayed in Figure 9 (for 1964) and Figure 10 (for 2009).

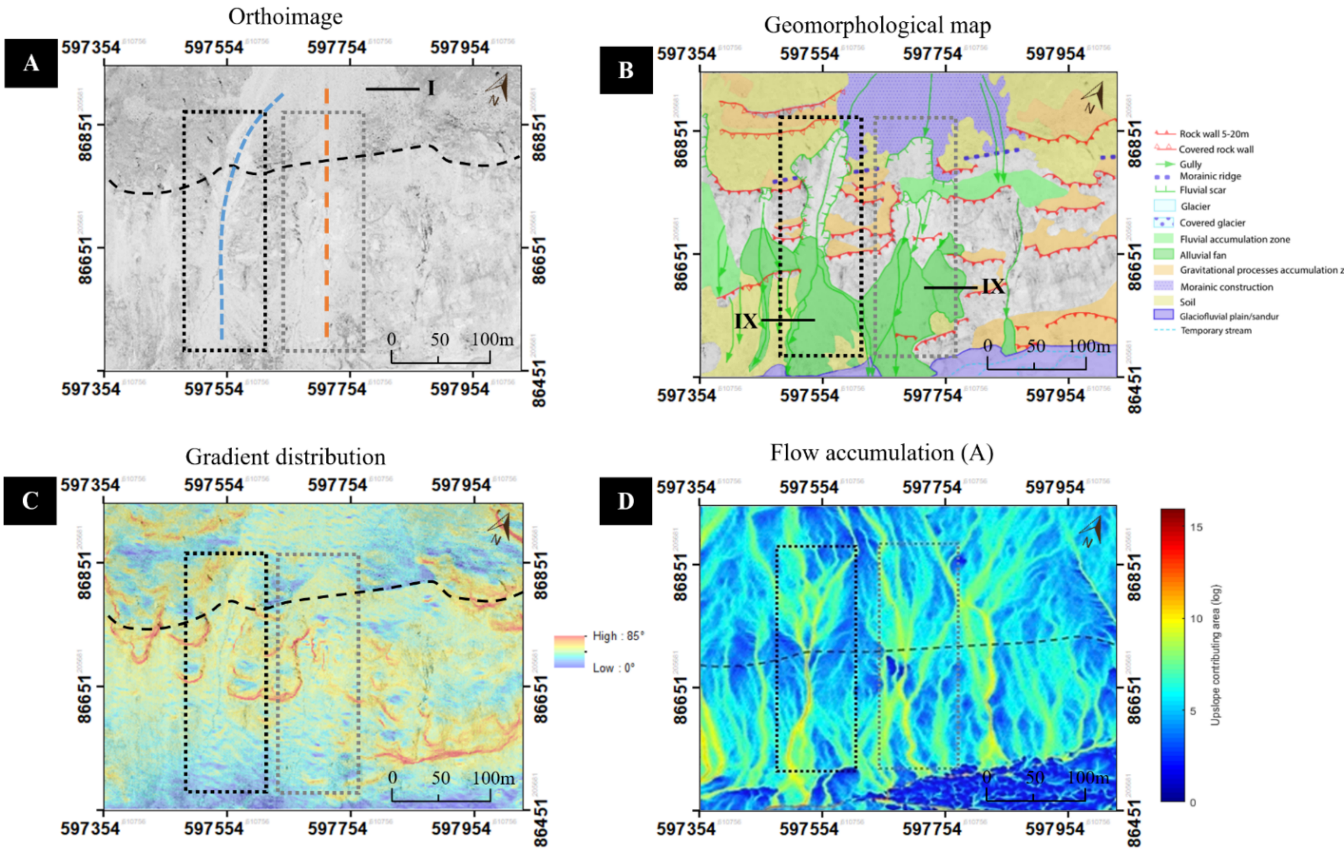

Figure 11: Geomorphic characteristics of the lateral moraine in 2009. A) orthoimage highlighting the location of the elevation profiles in Figure 10, B) geomorphologic map using the University of Lausanne classification (Lambiel et al., 2016), C) slope gradient and potential buffers, D) flow accumulation algorithm calculated with all pits filled with a threshold defined by the (Table 3). Dotted squares refer to SS location while dashed line represent the supposed LIA limit.

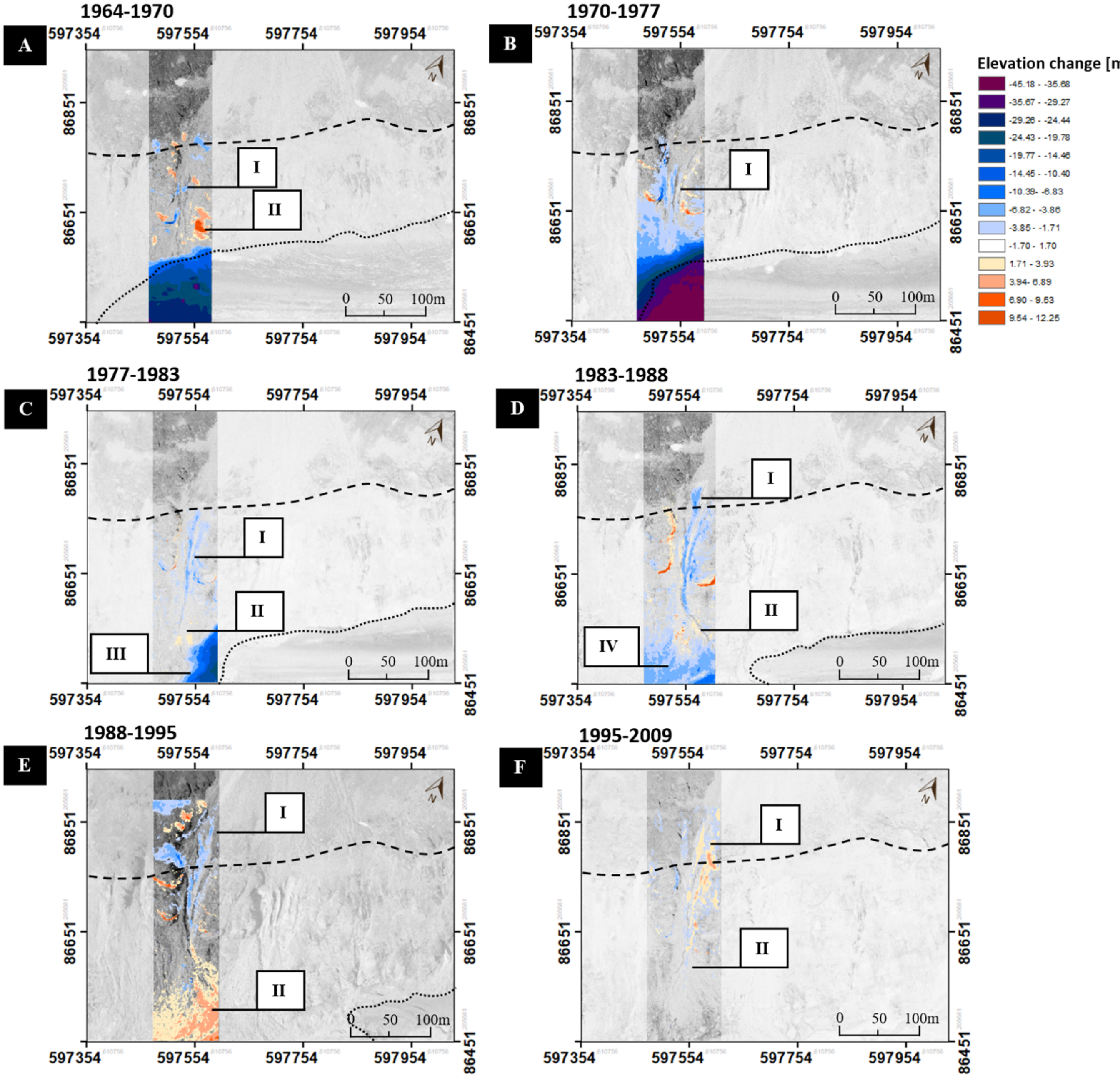

Figure 12: DEM of difference showing the historical geomorphological evolution of SS1. Dashed black line refers to the supposed LIA limit and the dotted one to the contact hillslope-glacier. Labels meaning: I = gully headward erosion, II = alluvial fan aggradation, III = valley floor exposure, IV = alluvial fan basal erosion.

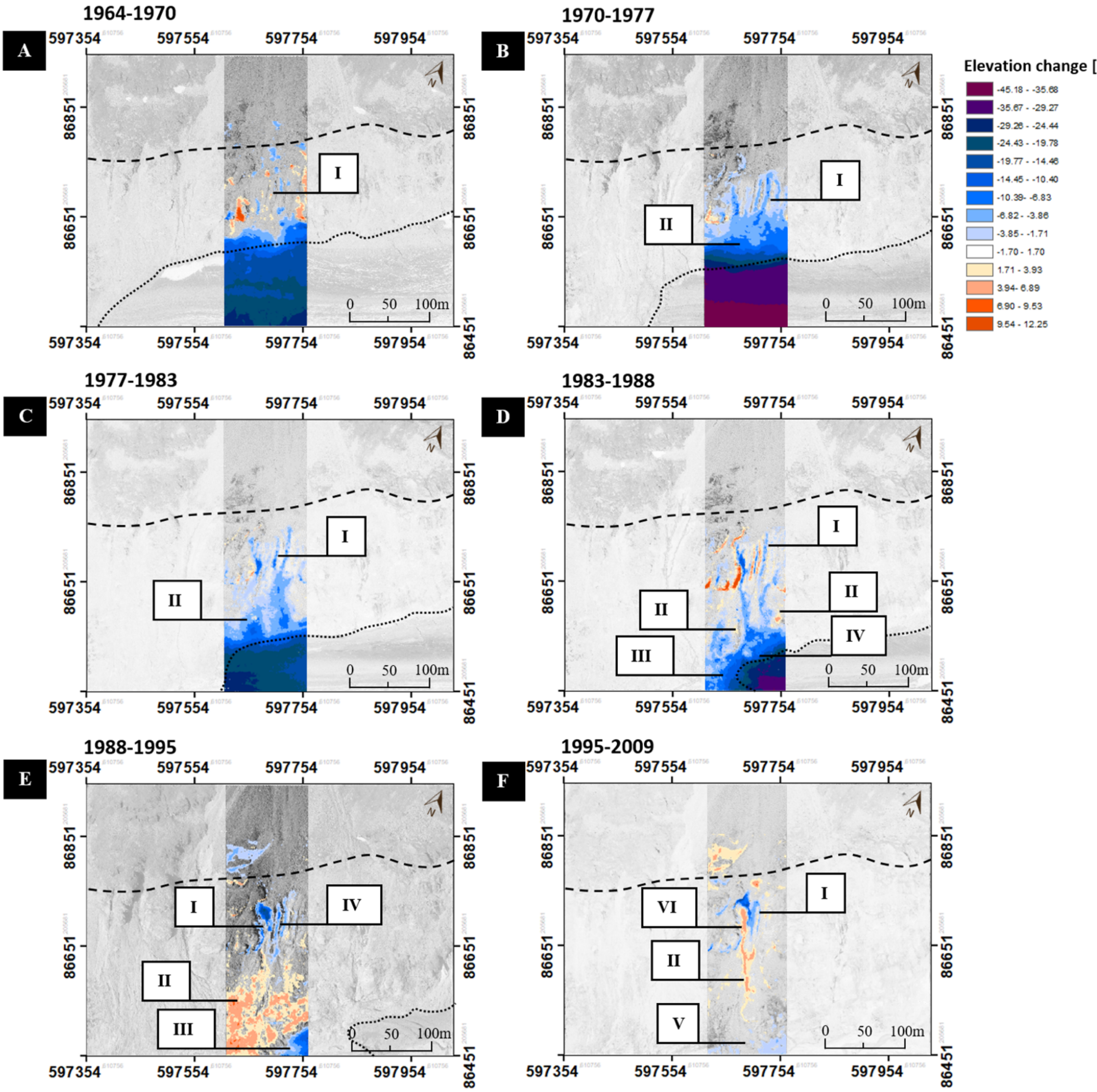

Figure 13: DEM of difference showing the historical geomorphological evolution of the SS1. Dashed black line refers to the supposed LIA limit and the dotted one to the contact hillslope-glacier. Labels meaning: I = gully headward erosion, II = alluvial fan aggradation, III = valley floor exposure, IV = bedrock patches exposure, V = alluvial fan basal erosion, VI = morainic material aggradation.

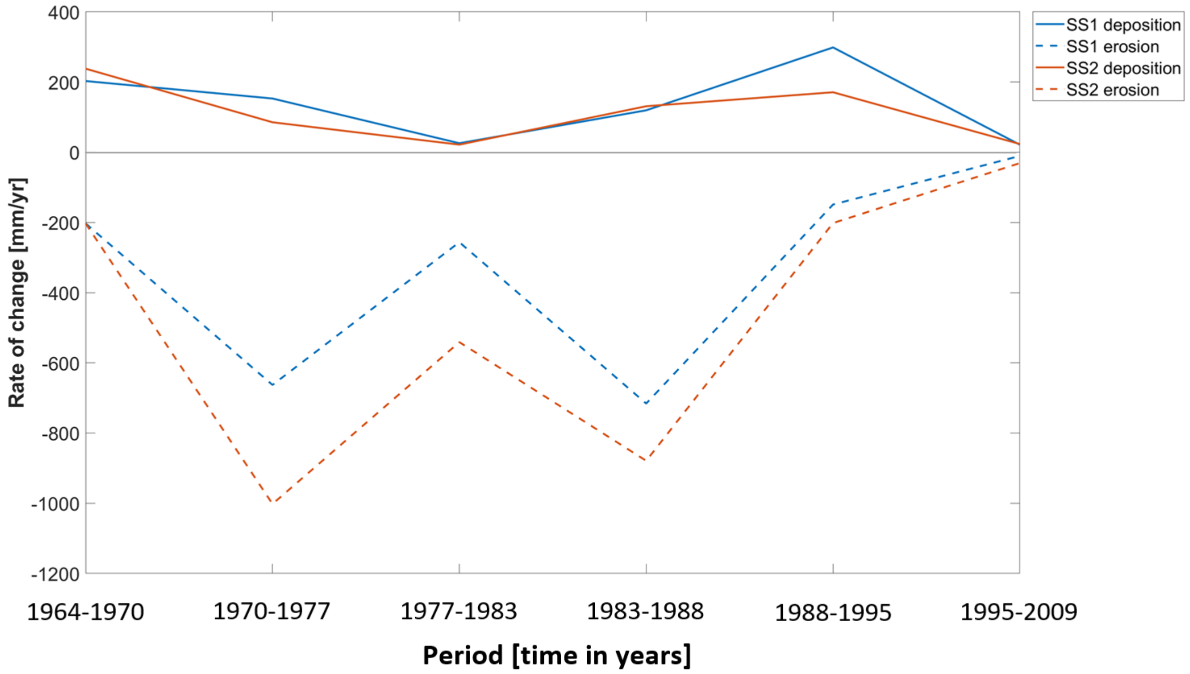

Figure 14: Quantification of total volume of remobilized sediment per period of investigation in both subsystems (values within the LoD have not been considered).

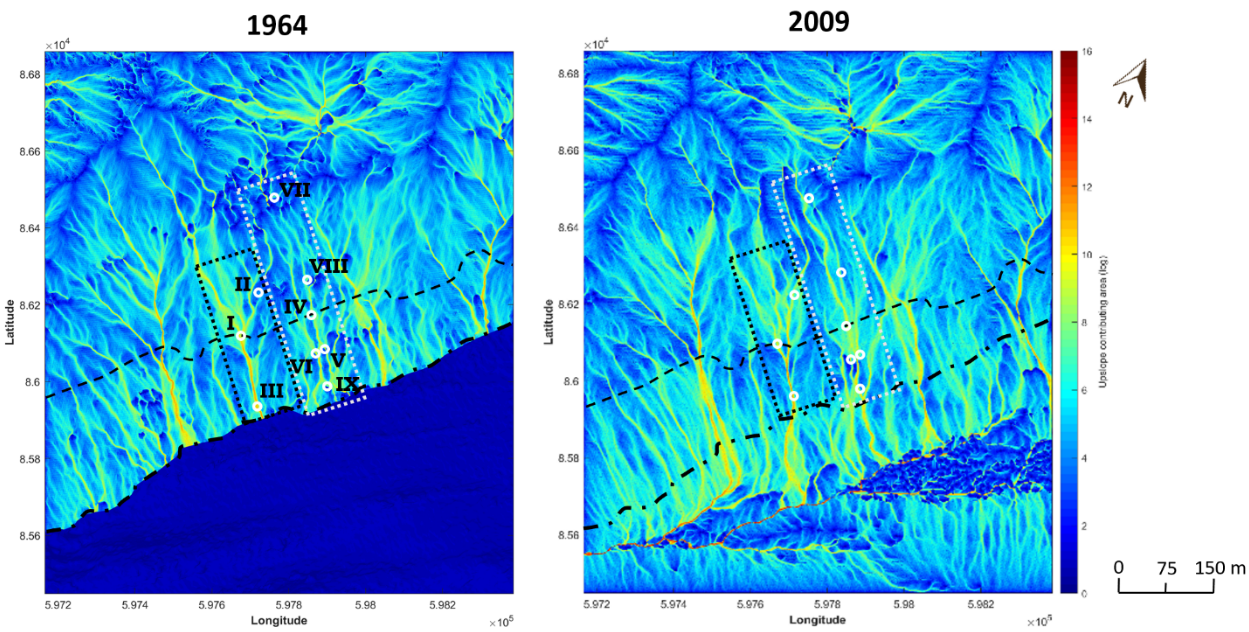

Figure 15: Upslope contributing area (Flowacc algorithm) for both 1964 and 2009 DEM with all pits filled. In the 1964 DEM, the glacier surface has been masked, and is not taken into account in computation as hydrological tools require impermeable ground (Carrivick et al., 2019). Numbers and circles refer to the potential buffers in figure 17; dotted lines define SS1 (black) and SS2 (gray) limits, dashed black line represent the supposed LIA limit and the dashed-dotted black line to the contact glacier-moraine. Hillslope extent is ca. 1.1 km².

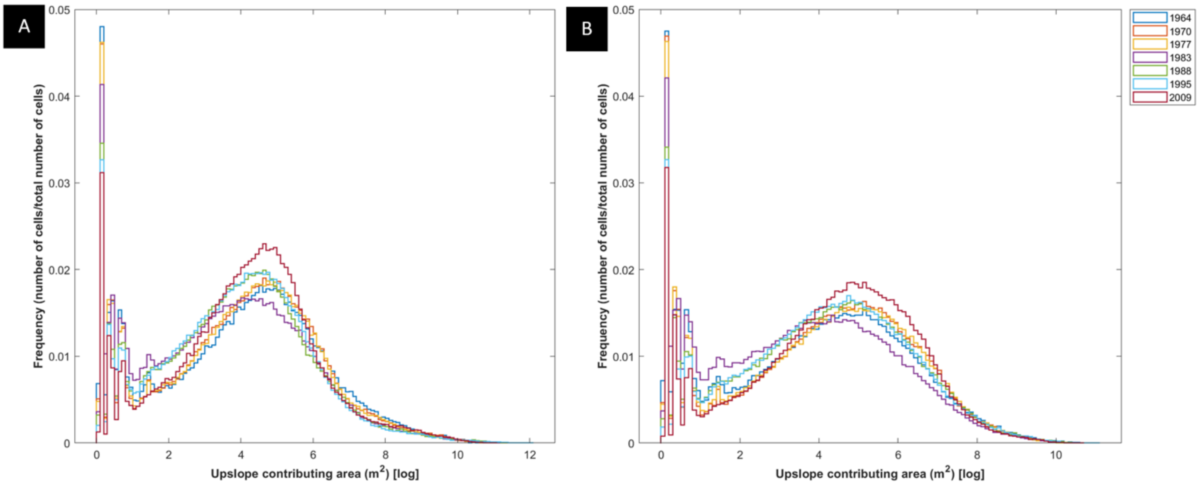

Figure 16: Histogram of flow accumulation area value for SS1 (A) and SS2 (B) in every year. Y-axis has been divided by the total number of pixels (246750 cells for SS1; 290745 cells for SS2).

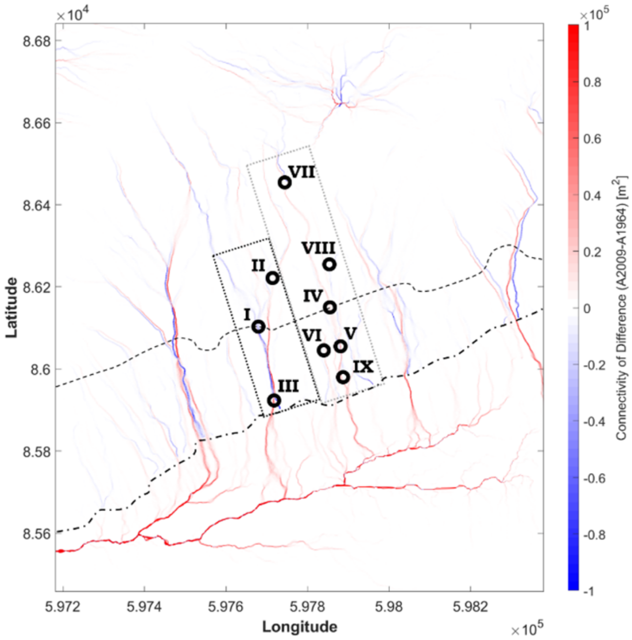

Figure 17: Changes in connectivity between 1964 and 2009. Supra-glacial streams have been masked, and are not taken into account in computation as the hydrological analyses require impermeable ground (Carrivick et al., 2019). Dotted lines define SS1 (black) and SS2 (gray) limits, dashed black line represents the supposed LIA limit and the dash-dotted black line to the contact glacier-moraine in 1964.

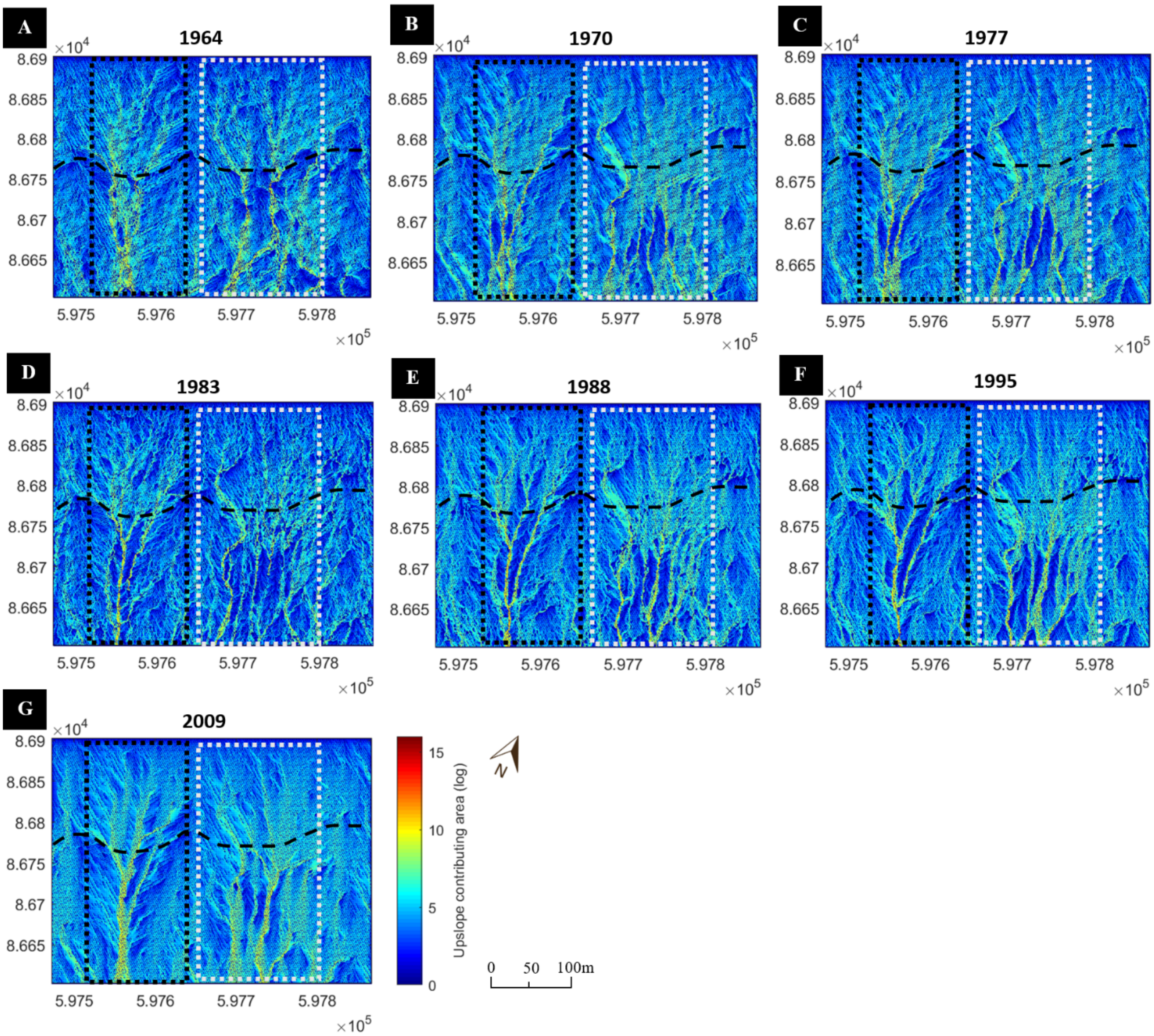

Figure 18: Evolution of upslope contributing area over time (histogram of cell values are represented in Figure 15). Dotted squares refers to SS1 and SS2 while dashed line highlight the location of the supposed LIA limit.

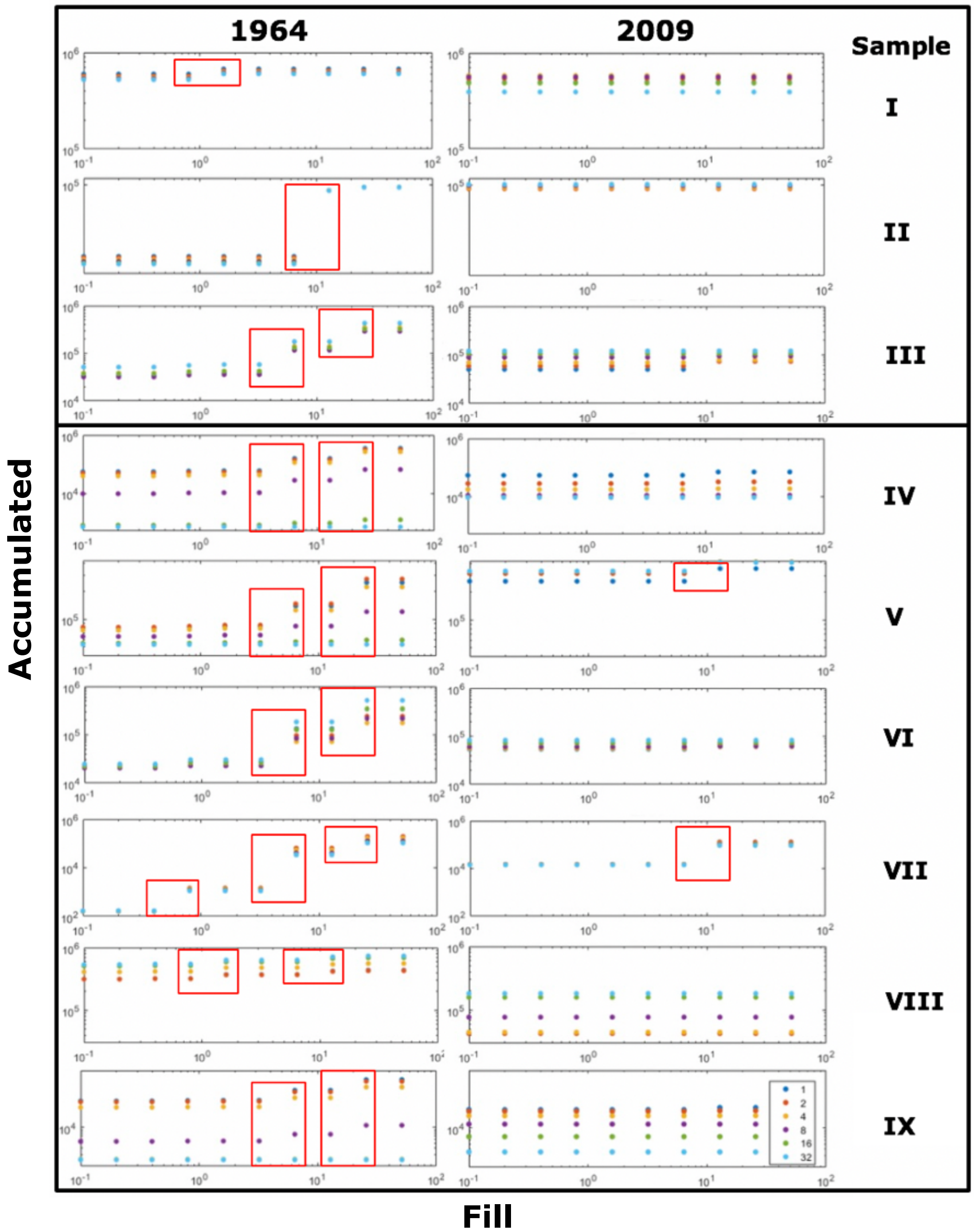

Figure 19 : Hydrological sediment connectivity quantification for 1964 and 2009 for selected regions highlighted in Figure 14. Point colors refer to the Holmgren flow routing parameters and the red squares to the increases in upslope contributing area.