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DOI: https://doi.org/10.1029/2021wr030221

Posted at the Zurich Open Repository and Archive, University of Zurich
ZORA URL: https://doi.org/10.5167/uzh-210685
Journal Article
Published Version

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Originally published at:
Maier, Fabian; van Meerveld, H J (2021). Long-term changes in runoff generation mechanisms for two proglacial areas in the Swiss Alps I: overland flow. Water Resources Research, 57(12):e2021WR030221. DOI: https://doi.org/10.1029/2021wr030221
Long-Term Changes in Runoff Generation Mechanisms for Two Proglacial Areas in the Swiss Alps I: Overland Flow

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Abstract In many areas of the world, the surface of the earth is changing rapidly. This can affect the partitioning of precipitation into overland flow (OF) and infiltration. OF can reach the stream quickly and thus strongly influences the streamflow response to precipitation. It can also cause surface erosion. However, our knowledge of the changes in OF responses during landscape evolution is still limited. To investigate how hillslope aging effects OF, we studied three plots on four different aged moraines (several decades to ~13.5 thousand years old) at a silicate and carbonate proglacial area in the Swiss Alps. We used sprinkling experiments to determine OF characteristics (such as the runoff ratio and timing) and used tracers (δ18H and NaCl) to identify the mixing of rainfall and soil water. Sediment concentrations and turbidity measurements provided an estimate of OF-driven soil erosion rates. The OF ratios were largest (42%) for the oldest moraines because the clay-rich layer at 20–40 cm below the surface caused saturated OF. However, OF occurred more frequently on the youngest moraines due to the high stone cover. Soil and vegetation development increase the soil water retention capacity and the pre-event water fractions in OF for the old moraines, but decreased the suspended sediment yield. The results show that OF characteristics and sediment transport change markedly during landscape evolution. This needs to be taken into account when simulating runoff and erosion responses for rapidly changing Alpine areas, and can—together with the outcomes for subsurface flow (see companion paper Maier et al., 2021, https://doi.org/10.1029/2021WR030223)—be used to improve landscape evolution models.

1. Introduction

Knowledge about the partitioning of rainfall into overland flow (OF; i.e., surface runoff), lateral subsurface flow (SSF), deep percolation, and soil water storage is important to understand and predict streamflow responses to rainfall events. OF also affects rainfall driven soil erosion (Huang et al., 2002; Poesen, 2017) and sediment transport (Gao, 2008; Prosser & Rustomji, 2004) and is thus an important factor for landscape evolution. Information on the partitioning of precipitation at the soil surface (i.e., into OF and infiltration) is therefore fundamental for hydrological models and landscape-evolution models (Gurtz et al., 2001; Jefferson et al., 2010).

There are two main processes that cause OF: infiltration-excess OF (i.e., Hortonian OF, HOF) and saturation-excess OF (SOF). HOF occurs when the rainfall intensity exceeds the infiltration capacity (Horton, 1933). HOF thus depends on the infiltration capacity of the soil surface and rainfall intensity. Higher rainfall intensities during natural rainfall events (Chaplot & Les Bissonnais, 2003; Ziegler et al., 2001) and sprinkling experiments (Lange et al., 2003; Markart, 1995; Mayerhofer et al., 2017; Römkins et al., 2002; Scherrer et al., 2007; Weiler & Naef, 2003) generally lead to a stronger HOF response. However, some studies report higher infiltration rates during high rainfall intensities, especially in highly heterogeneous soils with high macroporosity and a large spatial variability in the infiltration capacity (Langhans et al., 2011; Stone et al., 2008; Yu, 1999; Yu et al., 1997). HOF on sparsely vegetated hillslopes (Gomi et al., 2010) and unpaved roads (Ziegler et al., 2001) consisted mainly of event water (i.e., rainfall), regardless of the size of the event (Gomi et al., 2010), suggesting a lack of mixing with soil water (Ziegler et al., 2001). In contrast, SOF occurs when rainfall falls on a saturated soil. The volume of SOF thus depends on the rainfall volume and the available soil storage capacity (cf. the variable source area concept; Dunne & Black, 1970). Wetter antecedent conditions increase the occurrence of SOF (Bronstert & Bärdossy, 1999; Henninger et al., 1976; Mayerhofer et al., 2017). SOF tends to be a mixture of rainfall and exfiltrating soil water (Buttle, 1994), with the fraction of soil water in SOF decreasing with increasing rainfall intensity (Buttle, 1994; Sandström, 1996).
Biomat flow (Sidle et al., 2007) is another fast flow pathway near the soil surface. It is more rapid than SSF through the mineral soil but slower than OF. Biomat flow is important where hydrophobicity limits infiltration into the mineral soil (Gomi, Sidle, Miyata, et al., 2008; Gomi et al., 2010; Kim et al., 2014). However, biomat flow is difficult to measure. Most OF measurements in trenches and gutters include biomat flow as part of OF.

The amount of OF and the response of OF to rainfall depends on many surface characteristics (see Figure 1 for a simplified depiction of the main relations between soil and vegetation characteristics and OF). Vegetation growth increases the number of macropores and preferential flow pathways (Beven, 2018; Weiler, 2017), which increase infiltration and thereby decrease the occurrence of OF (arrows #6 and #10 in Figure 1). However, root exudates may cause water-repellency, reducing infiltration rates and can consequently enhance OF and biomat flow (Kim et al., 2014; Lemmnitz et al., 2008; Miyata et al., 2007; arrows #5 and #10 in Figure 1). Rocks at the surface also reduce surface infiltration (Arnau-Rosalén et al., 2008) and increase the susceptibility for OF if they are partially buried (arrow #8 in Figure 1), but their effect is different if they lay on the surface (Poesen & Bunte, 1996; Poesen & Ingelmo-Sanchez, 1992). Generally, larger rock fragments result in more concentrated flow onto the surrounding soil (Lavee & Poesen, 1991), which can either increase OF (Loos & Elsenbeer, 2011) or in cases with high microtopographic variation increase infiltration and reduce OF (Johnson et al., 1979; Streichen, 1984). Layers near the surface with a low hydraulic conductivity (e.g., clay layers) increase the likelihood of SOF (Elsenbeer, 2001; Lohse & Dietrich, 2005; arrows #1 and #4 in Figure 1), while soil compaction increases the likelihood of HOF (Mayerhofer et al., 2017; Zimmermann et al., 2006). Many studies have shown that the amount of OF is higher for steep terrain (El Kateb et al., 2013; Mu et al., 2015), but in other studies slope had no effect on the OF volume (Cerdà & García-Fayos, 1997; Yair & Klein, 1973; arrow #11 in Figure 1).

Although most of the above-mentioned vegetation, soil and surface characteristics can change dramatically over time, there are very few experimental studies that have investigated how the partitioning of rainfall at the soil surface changes during landscape evolution. As a result, our knowledge of how landscape evolution affects the quantity and characteristics of OF is still limited (Lohse & Dietrich, 2005), even though OF and related soil erosion and sediment transport processes are one of the main drivers of landscape evolution (Eppes et al., 2002; McAuliffe, 1994; Wells et al., 1987). Some studies have shown that the accumulation of clay in the B horizon of older volcanic soils (Lohse & Dietrich, 2005) and desert soils (Young et al., 2004) results in a lower saturated hydraulic conductivity ($K_{sat}$) and enhanced the occurrence of SOF. Onda et al. (2008) showed that after a severe forest fire, SSF was the dominant flow pathway because the loose ash layer promoted preferential flow and deeper percolation. However, after successive rainstorms compacted the ash and preferential flow paths became clogged, HOF became the dominant runoff mechanism. Contrary, in the Chicken Creek study in Germany, initial crusting led to lower infiltration rates and higher OF and erosion than expected based on the soil texture. But after vegetation was established, the development of macropores caused an increase in the $K_{sat}$ and the amount of OF was reduced (Holländer et al., 2009).

Alpine areas are rapidly changing (e.g., due to glacial retreat; Egli et al., 2010; Greinwald et al., 2021) and are characterized by high erosion rates and sediment transport (Blass et al., 2003; Geilhausen et al., 2013). However, experimental studies in Alpine areas have so far been limited (Scherrer et al., 2007). As a result, runoff generation processes, especially during extreme rainfall events, are still not well understood for high Alpine terrain and changes in runoff processes during landscape development have not been documented. We, therefore, conducted sprinkling experiments with different rainfall intensities on moraine chronosequences in two proglacial areas in the Swiss Alps with a different geology, to improve our understanding of how OF responses change during landscape evolution. More specifically, we asked the following research questions:

1. How do OF characteristics (runoff ratio, peak flow rate, timing, rainfall threshold, and event water fraction) change during the first millennia of landscape evolution on lateral moraines?
2. How are the changes in OF characteristics related to changes in soil- and vegetation characteristics?
3. How do the changes in the OF-, soil-, and vegetation characteristics affect event-based rainfall driven erosion rates?

We hypothesized that the increase in vegetation cover during landscape development and the associated increase in root length, root density, and macroporosity (Bundt et al., 2001; Johnson & Lehmann, 2016), increase the $K_{sat}$ of the soil (right side of Figure 1) and reduce OF and erosion (left side of Figure 1). We, furthermore, expected that soil and vegetation development, and in particular the increase in the clay and silt fraction during
Figure 1. Conceptual diagram that illustrates how the age of a hillslope influences hillslope characteristics (rectangles) and overland flow (OF) and soil erosion responses (ovals) due to the combined effects of time on vegetation and soil development, as described by the hypotheses. Soil and vegetation development affect the soil texture (in particular clay content and stone cover), soil aggregate stability, hydrophobicity, soil organic matter content, and root length and density. These properties, in turn, affect the saturated hydraulic conductivity ($K_{sat}$) and infiltration rates that determine the OF and biomat flow response and its characteristics (e.g., timing [$t_{peak}$], total volume [$Q_{total}$], runoff ratio [RR], peak flow rate [$Q_{peak}$], and event water fraction [$f_e$]). The soil characteristics and OF response affect the rainfall driven soil erosion rates. Other factors (such as the geology, topography, rainfall intensity, and amount, etc.) also affect OF generation and soil erosion. The arrows indicate the relations and feedbacks between the different hillslope characteristics and responses (solid for OF and dashed for soil erosion). The numbers next to the arrows refer to processes and relations described in the introduction and discussion sections. The main results of the manuscript are indicated by the figure numbers (squared brackets).

2. Study Sites

2.1. General Approach

In this study, we used a chronosequence (i.e., space-for-time) approach to study the effects of hillslope evolution (or soil and vegetation development) on OF and soil erosion. Because the rate of soil and vegetation development depends on the geology, e.g., the dissolution of carbonates is several orders of magnitude faster than silicate weathering (Binder, 2019; Jacobson et al., 2002; Lasaga, 1984; White & Buss, 2003; Williamson & Rims tidt, 1994), we selected chronosequences in two geologically different areas in the Swiss Alps: (a) the Sustenpass area on silicate bedrock and (b) the Klausenpass area on carbonate bedrock (Table 1 and Figure 2). At each study area, we studied four moraines of different ages and installed three bounded runoff plots on each moraine. We took the same measurements and conducted the same sprinkling (i.e., rainfall simulation) experiments on each of the 24 plots. The measurements at the Sustenpass were taken between July and September 2018, while the data at the Klausenpass were obtained between August and September 2019.

Sprinkling experiments are very useful to study rainfall-runoff processes during a range of rainfall intensities, including extreme events (Lange et al., 2003; Scherrer et al., 2007; van Meerveld et al., 2014). Chronosequences are fundamental for understanding changes in hillslope functioning over time and the feedback mechanisms between the myriad of ecological, hydrological, and geomorphological processes (as indicated by the many arrows in Figure 1). However, so far, sprinkling experiments have rarely been coupled with a chronosequence approach to identify changes in runoff processes during landscape development (Bernasconi et al., 2011; Rosenqvist et al., 2010).

The chronosequence approach used in this study assumes that only the time since soil and vegetation development started varies between the study moraines and that all other factors are similar or accounted for. Thus, we assume that the main differences between the four moraines in a study area are caused by time (i.e., the age of the moraines). Time is one of the five dominant soil formation factors (Jenny, 1941). The other four factors that affect soil development (Jenny, 1941) are either assumed to be similar within a study area (parent material and climate) or the variation is accounted for in the study design (topography and biota).

Although the chronosequence approach has its limitations (e.g., Johnson & Miyanishi, 2008; Pickett, 1989; Walker et al., 2010), it is one of the few feasible field methods to study the factors that influence soil or vegetation development. It has previously been used to study soil formation (e.g., D’Amico et al., 2014; Kiczka et al., 2011; Stevens & Walker, 1970), the effects of land use change on soil hydrological characteristics (e.g., Nyberg et al., 2012; van Meerveld et al., 2021; Zimmermann et al., 2006), in ecology (e.g., Aide et al., 2000; Richardson et al., 2005; Walker et al., 2010), and has been used in several studies in the Alps (e.g., Bernasconi et al., 2011; Burga et al., 2010; D’Amico et al., 2014; Dümig et al., 2011; Temme & Lange, 2014).

2.2. The Two Study Areas

The Sustenpass study area is located in the foreland of the Steingletscher, south of the Sustenpass mountain road (47°43’N, 8°25’E) in the internal Alpine massif. The bedrock of the highly polymetamorphic
| Table 1 | The Main Characteristics of the Two Study Areas and the Studied Moraines |
|---------|-------------------------------------------------|
| Geology | Sustenpass                                       | Klausenpass                                  |
| Moraine age (yr) | 10,000 | 3,000 | 160 | 30 | 13,500 | 4,900 | 160 | 80 |
| Elevation (m a.s.l.) | 1873–1882 | 1888–1914 | 1981–1989 | 1952–1959 | 2001–2017 | 2016–2019 | 2025–2038 | 2036–2042 |
| Slope (°) | 18–29 (24) | 25–32 (29) | 25–37 (29) | 22–33 (26) | 33–38 (35) | 28–34 (32) | 28–44 (36) | 29–35 (32) |
| Aspect | NE | S | NE | NE | NE | NE | NE | NE |
| Tortuosity index | 0.64–1.14 (0.95) | 0.33–0.65 (0.52) | 0.29–0.40 (0.32) | 0.36–0.55 (0.48) | 0.25–0.30 (0.28) | 0.26–0.68 (0.40) | 0.27–0.36 (0.32) | 0.27–0.85 (0.66) |
| Vegetation cover (%) | 90–100 (93) | 60–85 (72) | 80–95 (85) | 30–50 (42) | 85–95 (90) | 74–82 (78) | 33–39 (36) | 17–27 (22) |
| Hydrophobicity | 2.46–9.09 (5.25) | 2.04–3.34 (2.83) | 1.40–2.55 (1.94) | 0.82–1.60 (1.22) | 5.85–14.51 (9.36) | 2.96–6.73 (5.05) | 1.04–2.46 (1.76) | 0.91–2.30 (1.72) |
| Dominant species | Rhododendron ferrugineum; Vaccinium myrtillus; Calluna vulgaris | Calluna vulgaris | Calluna vulgaris | Calluna vulgaris | Calluna vulgaris | Calluna vulgaris | Calluna vulgaris | Calluna vulgaris |
| Soil texture | Silty-loam | Sandy-loam | Sandy-loam | Sandy-loam | Silty-loam | Sandy-loam | Sandy-loam | Loamy-sand |
| Soil type | Dystric Cambisol | Skeletal Cambisol | Hyperskeletal Leptosol | Hyperskeletal Leptosol | Calcaric Skeletal Cambisol | Calcaric Skeletal Cambisol | Calcaric Skeletal Cambisol | Orthoskeletal Leptosol |

Note. Ranges represent the range in average values for the three plots per moraine, while the value in parentheses represents the average for the three plots per moraine. Multiple expositions for a moraine indicate that plots were located on both sides of the moraine.

*Calculated based on the normalized line length for ten measurements with a 1.5 m long microtopography profiler (cf. Leatherman, 1987) per plot using the method of Bertuzzi et al. (1990). *Calculated based on the ratio of the sorptivity for 95% ethanol and water (Hallett et al., 2001; Lichner et al., 2007). Soil type and texture were defined for each moraine by Musso et al. (2019) based on the World Reference Base for Soil Resources (IUSS Working Group WRB, 2015).
“Erstfelder”-gneiss-zone consists mainly of pre-mesozoic metagranitoids, amphibolites, and biotite-plagioclase-gneiss (Labhart, 1977). The Klausenpass study area is located in the proglacial area of the Griessfirn glacier, south of the Klausenpass mountain road (46°50ʹN, 8°49ʹE) in an area where the Helvetic nappes have been thrust over the Aarmassif and the Infra-helvetic complex. In the Klausenpass study area, the thrust forms the contact between the Permo-Triassic and Jurassic and Cretaceous limestones and Paleogene flysch and molasse (Frey, 1965; Pfiffner, 2014). The proglacial area is dominated by calcareous material (Globigerinen limestone and Globigerinen marl), schist (Pectinien schist), and some quartzites.

Both study areas are situated above the tree line and are characterized by an Alpine climate. At the Grimsel Hospiz, the closest long-term weather station at a comparable elevation (1979 m.a.s.l.; located 20 and 50 km from Sustenpass and Klausenpass, respectively), the average monthly temperature varies between 10.0°C in July and −5.4°C in February; the mean annual air temperature is 1.9°C. Mean annual precipitation at the Grimsel Hospiz is 1,856 mm yr⁻¹ and is evenly distributed during the year (1981–2010 period; MeteoSwiss, 2016). Snowfall events are common during the whole year (MeteoSwiss, 2016). Although the hydrological regime is dominated by snowmelt, high intensity rainfall events in summer are important as well (Umbricht et al., 2013).

2.3. Selection and Age Dating of the Moraines

At each study area, there are well-preserved moraine remnants beyond the little ice age extend (King, 1974). Schimmelpfennig et al. (2014) created a detailed and comprehensive chronology of the moraines at the Sustenpass study area based on beryllium-10 dating. Based on her work, the glacier extent maps for the last 150 yr, and aerial images (Swiss Federal Office of Topography—“Journey through time”), four moraines were selected (Figure 2a). The two older moraines are 10 and 2–3 thousand years in age and are hereafter referred to as the
10-ky moraine and 3-ky moraine. The two younger moraines are approximately 160 and 30 yr in age (hereafter called the 160-y and 30-yr moraines, respectively). At the Klausenpass study area, we also selected four moraines (Figure 2b). The two older moraines were dated with the radiocarbon method by Musso et al. (2019) to 13.5 thousand and 4.9 thousand years of age and are hereafter referred to as the 13.5-ky and 4.9-ky moraine. The second youngest moraine is a remnant of the little ice age and was dated based on historical maps and aerial images. It is hereafter referred to as the 160-y moraine. The youngest moraine was dated based on year ring analysis of Dryas octopetala and Salix retusa because the glacier extent maps were not very clear. It is ∼60–100 yr in age and is hereafter referred to as the 80-yr moraine. Although the moraines at the Sustenpass and Klausenpass study areas differ in age, we consider them to be of a similar age class, that is, the 10-ky moraine at Sustenpass and the 13.5-ky moraine at Klausenpass are the oldest moraine class, the 3-ky moraine at Sustenpass and the 4.9-ky moraine at Klausenpass the second oldest moraine class, the 160-y moraines at both study areas form the second youngest moraine class, and the 30-yr moraine at Sustenpass and 80-yr moraine at Klausenpass belong to the youngest moraine class.

All of the moraines are lateral moraines (e.g., Musso et al., 2020; Schimmelpfennig et al., 2014). We, therefore, assume that the formation process, erosive forces, and depositional settings are similar. Because the moraines in each study area were formed by the same glacier and the geology in each study area is relatively uniform, we also assume that the parent material for each of the four moraines in a study area was similar. The mineralogy of the moraines at each study area is indeed very similar (Musso et al., 2021). The four moraines at each study area are located within 400 m from each other and thus, have a similar climate as well.

### 2.4. Moraine Characteristics

#### 2.4.1. Vegetation

The vegetation cover increased with the age of the moraines for both study areas (Greinwald et al., 2021), but at the Klausenpass, vegetation cover mainly differed between the two oldest and youngest moraines (Table 1, Figure 3). The surface exposed rock area varied between 11% and 45% for the youngest moraines (Figure 3). Rocks were present on the older moraines as well, but surface cover could not be determined due to the high vegetation density (Figure 3).

The oldest moraines were covered by Alpine rose (Rhododendron ferrugineum). The Alpine rose was much more dominant on the oldest moraine at the Sustenpass than at the Klausenpass (70% vs. 10% average cover). The oldest moraine at the Klausenpass, the second oldest moraines at both areas (3-ky and 4.9-ky), and the 160-y moraine at Sustenpass are Alpine grasslands that are covered by sedges and small shrubs (Table 1). Vegetation cover on the 160-y moraine at the Sustenpass was higher than for the 3-ky moraine (Figure 3), where scree fields were more abundant due to rockfall.

The surface of the 160-y moraine at Klausenpass and the youngest moraines at both study areas consist mainly of large scree fields and bare soil (Figure 3). The vegetation was characterized by pioneer species and early successional communities (Table 1) that are typical for coarse-textured soils (Hudek et al., 2017; Jonasson & Callaghan, 1992; Lichtenegger, 1996; Pohl et al., 2011). These species have shallow root systems.

All moraines are occasionally affected by low intensity grazing by sheep or cows. However, we assume that this did not have a significant effect on the pedological, biological and hydrological development of the hillslopes or the OF and erosion responses.

#### 2.4.2. Topography

All moraines are very steep, but the average slope was a bit higher for the moraines at the Klausenpass study area than at the Sustenpass (Table 1). The surface topography was most variable for the 10-ky moraine at the Sustenpass, (i.e., highest Tortuosity Index; cf. Bertuzzi et al., 1990) due to the hummocky character of the abundant Alpine rose. Contrary, at the Klausenpass, the surface of the oldest moraines was relatively smooth and the Tortuosity Index was highest for the youngest moraine due to the presence of large rocks (Figure 3 and Table 1).

#### 2.4.3. Soil Type and Texture

Soil type and texture were defined for each moraine by Musso et al. (2019) based on the World Reference Base for Soil Resources (IUSS Working Group WRB, 2015). At Sustenpass, the soils were classified as Dystric Cambisols
Water Resources Research

(10-ky), Skeletic Cambisols (3-ky), and Hyperskeletic Leptosols (160-y and 30-y). At Klausenpass, the soil classes are Calcaric Skeletic Cambisols (13.5 and 4.9-ky), Hyperskeletic Leptosols (160-y), and Orthoskeletic Leptosols (80-y; Table 1).

The soils of the older moraines had a finer texture (silty-loamy soil texture for the older moraines vs. a gravelly and sandy texture for the younger moraines; Table 1) and were more developed. The oldest moraines also had a well-developed litter and organic layer. However, the organic litter layer at the surface and clay-rich layers at 20–40 cm below the surface were more pronounced for the oldest moraines at the Sustenpass than at Klausenpass. The rooting density and depth were also higher for the older moraines than the younger moraines, which foster heterogeneous macropore flow and finger flow (Hartmann, Semenova, Weiler, & Blume, 2020). Particularly at the Klausenpass, there was a very dense 5 cm thick root mat on the 13.5- and 4.9-ky moraine that facilitated biomat flow. In contrast, the coarse material on the young moraines enables homogeneous gravity driven matrix flow (Hartmann, Semenova, Weiler, & Blume, 2020). For a more detailed description of the soil characteristics of the study moraines, see Maier et al. (2021; companion paper).

2.5. Study Plot Selection

We assessed the structural vegetation complexity for ten 1 m × 1 m subplots on each moraine based on the vegetation cover, number of different species, and functional diversity (based on stem growth form, root type, clonal growth organ, seed mass, Raunkiaer’s life form, leaf dry matter content, nitrogen content, and specific leaf area [Garnier et al., 2016; Greinwald et al., 2021]). We selected the area with the lowest, intermediate, and highest structural vegetation complexity for the establishment of the bounded runoff plots to cover as much of the potential variability within each moraine as possible (Figure 3).

The 4 m × 6 m runoff plots were relatively planar (no extreme convergence or divergence) and were established in the direction of the main slope of the moraine. The distance between the three plots on a moraine was less than 50 m (Figures S1 and S2). All plots were located at a midslope location or closer to the ridge, except for the low
and high complexity plots at the 30-y moraine at Sustenpass, the high complexity plot of the 3-ky moraine at the Sustenpass, and the high complexity plot of the 160-y moraine at the Klausenpass, which were located closer to the bottom of the slopes (Figure S1).

3. Methods

3.1. Soil Surface Characteristics

3.1.1. Saturated Hydraulic Conductivity ($K_{sat}$)

The vertical saturated hydraulic conductivity $K_{sat}$ of the surface was measured at three locations next to each plot using a Double Ring Infiltrometer with an inner diameter of 20 cm, according to the ASTM D3385-03 standard test method. To facilitate vertical insertion of the infiltrometers by at least 5 cm on the steep slopes, the lower end was cut at an angle of 20° and 30°, which corresponds to the median and average slope of the moraines, respectively. The depth of ponding varied between 15 and 25 cm. The $K_{sat}$ was based on the steady state infiltration rate. The $K_{sat}$ of the subsurface (5–20 and 20–40 cm) was measured at the same locations with a constant head permeameter (Amoozegar, 1992, 1989; see Maier et al., 2021, for more details). We report the median and average value of the nine measurements at each depth per moraine. We acknowledge that for a robust estimation of the average or median $K_{sat}$ value many more measurements are needed (see, e.g., Harden & Scruggs, 2003; Zimmermann, 2008) but this was not possible due to time limitations. However, some of the expected variability and trends are already revealed by the nine measurements per moraine.

3.1.2. Hydrophobicity

We used a Mini Disk Infiltrometer (Decagon Devices, Inc., Pullman, WA, USA) to determine the hydrophobicity of the soil surface below the most dominant and second most dominant species at each plot. We carefully removed the organic material (lichens, mosses, leaves, and litter) with a trowel prior to the measurements. The Mini Disk Infiltrometer was filled with either water or a 95% ethanol solution and infiltration volumes were recorded until a steady-flow rate was achieved. We used the index of soil water repellency, $R$, which is based on the ratio of the sorptivity for the 95% ethanol and water measurements (Hallett et al., 2001; Lichner et al., 2007) to estimate how much sorptivity is reduced by the water repellency of the soils. Values of $R$ larger than 1.95 indicate sub-critically water repellent soils (Tillmann et al., 1989). The average of the two measurements was used to represent the hydrophobicity for each plot.

3.1.3. Soil Aggregate Stability

Six soil samples were collected at 0–10 cm below the surface of each plot with a Humax probe to determine the soil aggregate stability according to the method for stone-rich soils of Bast et al. (2015). This method is based on the wet-sieving method using a 20 mm sieve. An aggregate stability coefficient (ASC) value of 1 indicates a completely stable soil sample, whereas a value of 0 indicates complete dispersion. The median ASC-value for the six cores was used to determine the aggregate stability for each plot. The median values for all 18 measurements per moraine were used to represent the ASC for each moraine.

3.2. Overland Flow Measurements

3.2.1. Plot Setup

The 4 m × 6 m runoff plots were bounded using plastic sheeting (Figure 4a) that was inserted ~5 cm into the ground to minimize lateral in- and outflow of OF. Four soil moisture sensors (Trübner SMT100 and Decagon 5 TE) were installed at 10 cm below the surface on both sides of the plot near the location of the upper and lower boundaries of the plot. At the bottom of the plot, a trench was excavated and pond foil was inserted 2–5 cm into the ground at 2–10 cm below the surface and connected to a gutter (Figure 4b). Thus, the gutter did not only collect OF but also biomat flow and very shallow SSF. The gutter was covered by plastic panels so that rainfall could not enter the gutter (Figure 4a).

During the sprinkling experiments (described below) water flowed from the gutter via a 5 cm diameter hose to an Upwelling Bernoulli Tube (Stewart et al., 2015), which contained a pressure transducer, electrical conductivity (EC), and temperature sensor (DCX-22-CTD, Keller). Another pressure transducer measured the barometric
pressure. This allowed us to determine the flow rate, EC and temperature of the OF at a 1 min interval. A turbidity sensor (Cyclops-7, Turner), installed in the hose between the gutter and the Upwelling Bernoulli Tube, recorded the turbidity (in NTU) at a 1-min-interval.

During natural rainfall events, the hose from the OF gutter of the plots with the lowest vegetation cover drained into a 200 l closed barrel with a pressure transducer (DCX-22-CTD, Keller). The relation between the water level and volume of water in the barrel was used to determine the rate of OF during the natural rainfall events. The best OF data at the Klausenpass were collected between July and August 2019. Unfortunately, no reliable OF data are available for natural rainfall events at the Sustenpass study area.

3.2.2. Sprinkling Experiments

We conducted three sprinkling experiments on each plot. Three adjustable sprinklers (Senninger I-Wob; nozzle number 22) were installed along a transect through the middle of each plot at 2 m above the soil surface (Figure 4a). The sprinklers were connected via garden hoses to a water meter and a fire hose that drained two 4 m³ water reservoirs filled with water from a nearby stream. The reservoirs were located at least ∼20 m above the plots to guarantee a sufficient pressure for the sprinklers. The sprinklers applied rainfall to an area that extended at least 0.5 m upslope from the plot and 6 m to the left and right of the plot (as well as several meters below the plot).

To obtain the different rainfall intensities, either one, two, or all three sprinklers were used to apply water to the plots. The aim was to irrigate each plot with an intensity that represents a 2.3-, 30- and 100-yr return period event (reference period: 1966–2015; MeteoSwiss, 2017). The events are referred to as LI, MI, and HI (low, medium, and high intensity) experiments in the remainder of the text. The three sprinkling experiments on each plot were conducted on three consecutive days, with the lowest intensity experiment on the first day and the highest intensity experiment on the third day. There were at least 24 hr without precipitation prior to the start of the LI sprinkling experiment. The plots were covered with big tarps before and in between the experiments in case natural rainfall was expected to occur during this period. Most of the drainage of the topsoil occurs within 24 hr and for the coarse soils on the young moraines this may be even faster. We, therefore, assume that the soils were close to field capacity for the MI and HI sprinkling experiments. Waiting longer was not possible due to the frequent rainfall events and short snow free season.
The rainfall intensities were measured using two tipping bucket rain gauges (RK400-04) and four manual funnel gauges installed on and next to each plot (Figure 4a). The rainfall intensity for the LI, MI, and HI experiments varied from plot to plot (Table S1) due to wind drift and differences in the pressure at the sprinkler heads. The average and standard deviation of the intensities were 20 ± 5, 44 ± 13, and 62 ± 16 mm hr⁻¹ for the LI, MI, and HI experiments, respectively. Each experiment continued until the total volume of water that had infiltrated (i.e., applied rainfall minus OF) exceeded 20 mm. The typical sprinkling durations were ∼75, 45, and 30 min, for the LI, MI, and HI experiments, respectively (see Table S1 for details). The spatial uniformity of the applied rainfall was assessed for each experiment based on the rainfall measurements. The median coefficients of variation for the measured rainfall amounts was 34% (±16%).

The median drop size (determined with the oil method; cf. Eigel & Moore, 1983) at a rainfall intensity of 39 mm hr⁻¹ was 1.15 ± 0.4 mm (range: 0.44–2.7 mm). This corresponds well with the median drop size of ∼1.1 mm for mountain regions in Western Europe (Hachani et al., 2017) and the median drop size of 0.1–1.5 mm for orographic precipitation (Blanchard, 1953). Therefore, we assume that the kinetic energy of the applied rainfall was somewhat comparable to that of natural rainfall.

3.2.3. Water Sampling and Analyses

For the LI experiments, the 8 m³ of stream water in the reservoirs was enriched with 400 ml of 99.8% Deuteriumoxid. For the MI and HI experiments, the water was enriched with 200 and 100 ml of Deuteriumoxid, respectively. For the HI experiments, 2 kg NaCl was added to the water as well. The water in the reservoirs was stirred thoroughly for at least 10 min before turning the sprinklers on.

Rainfall samples were collected from the funnel rain gauges after each sprinkling experiment. Soil water samples were taken before the first (LI) experiment using suction lysimeters (SMF-30, UMS) installed at ~30 cm below the soil surface. The isotopic composition of the soil water was also determined for soil samples taken at 10 cm intervals from the top 30 cm of the soil at two locations next to each plot (bag method; cf. Jiménez-Rodríguez et al., 2019). The EC of the rainfall and soil water samples was measured in the field using a Multi 3420 conductivity probe (WTW Measurement Systems Inc).

Samples of the OF were taken from the outflow of the Upwelling Bernoulli Tubes using 500 ml wide-mouth polyethylene bottles. Sampling started as soon as flow started and continued until outflow stopped. The sampling frequency was higher at the beginning of the experiments because higher rates of change were expected during the rising limb of the hydrograph (cf. Zhao & Hou, 2018). The samples were filtered using 1.6 μm filters (GF/A Whatman) to remove the sediment. These filters were oven dried and weighted before and after use to determine the total amount of sediment in each sample, and thus the sediment concentration for each sample.

The rainfall and soil water samples and the filtered OF samples were filtered using 0.45 μm syringe filters (SimplePureTM Syringe Filter) and stored in 20 ml glass vials without any air bubbles or headspace at 4°C. The samples were analyzed for stable water isotope analysis using a Cavity Ring-Down Spectrometer (L2130-I Picarro Inc.) at the isotope laboratory of the Chairs of Hydrology at the University of Freiburg (Germany). We use the delta notation relative to Vienna Standard Mean Ocean Water. The precision of the analysis was ±0.05‰ for δ¹⁸O and ±0.35‰ for δ²H.

3.3. Data Analyses

3.3.1. Overland Flow Response Characteristics

We represent the soil wetness conditions prior to the start of the sprinkling experiments and the rainfall events using the Antecedent Soil-moisture Index (ASI), which is the product of the average volumetric soil water content (θ) and the depth (10 cm) of the soil moisture sensors (Haga et al., 2005; Penna et al., 2015). To be able to compare the OF rates of the different experiments, we scaled the OF rates by the average rainfall intensity (Pᵢ) for each experiment (Table S1). To account for the small differences in the EC of the applied rainfall for the HI experiments, the EC of the OF was similarly scaled by dividing it by the EC of the rainfall (scaled EC).

For each experiment that resulted in OF, we determined the time until OF occurred (tlag), the total amount of rainfall before OF occurred (Platex), the peak flow rate (Qpeak), the total OF volume (Qtotal), the ratio of the total OF
volume and total precipitation (runoff ratio), and the total drainage after the experiment had ended \((D)\). The total volume of \(DF\) was converted to a depth using the field-measured plot areas. For the drainage after the experiment ended \((D)\), we determined the total amount of \(DF\) that drained from the plots after the sprinklers had been turned off for 5 min, which we regard as the maximum time for pure \(DF\) to reach the gutter. We, furthermore, calculated the effective infiltration rate for each experiments based on the difference between the peak flow rate and the average rainfall intensity \((I = P_{\text{avr}} - Q_{\text{peak}})\). Note that this represents the minimum infiltration rate during the experiment and is an average for the entire plot. For a visual description of these hydrograph characteristics, see Figure S3.

### 3.3.2. Hydrograph Separation

We assessed the fraction of the rainfall (event water fraction, \(f_e\)) and soil water in \(DF\) for the LI experiments based on a two-component hydrograph separation using \(\delta^2H\) as the tracer (cf. Sklash et al., 1976). We used the average \(\delta^2H\) value of the rain samples collected during the experiment to represent the rainfall, and similarly the average \(\delta^2H\) of all soil water samples (from the suction lysimeter and the soil cores) to characterize the soil water isotopic composition.

For the MI and HI sprinkling experiments, we performed a three-component hydrograph separation (cf. the method of Gibson et al., 2000) using \(\delta^{18}O\) and \(\delta^2H\) as tracers to distinguish between event water (rainfall applied during the experiment), previous event water (i.e., rainfall applied during the LI experiment for the MI experiment and the rainfall applied during the MI experiment for the HI experiment), and soil water (collected prior to the LI experiment). For the two outliers for the MI experiment on the 10-ky medium complexity plot and the two outliers for the HI experiment on the 160-ky moraine low complexity plot at Sustenpass, we used a least-distance-approach to project them into the mixing space. A four-component mixing analysis for the HI experiments was not attempted because the visual inspection of the mixing spaces of EC, \(\delta^{18}O\), and \(\delta^2H\) (Figure S4) did not suggest a notable contribution of the rainfall applied during the LI experiments to \(DF\) during the HI experiments.

We used the method of Genereux (1998) to determine the uncertainty for the hydrograph separation calculations. For the uncertainty of the event water and previous event water compositions, we used the standard deviation of the rainfall samples collected during the experiment \((n = 2–4)\). For the uncertainty of the soil water composition, we used the standard deviation of all soil water samples collected prior to the first sprinkling experiment at each plot \((n = 5–12)\). For the uncertainty of the isotopic composition of the \(DF\), we used two times the precision of the measurements.

### 3.3.3. Sediment Yield

The suspended sediment concentrations (from the filtered samples) were correlated to the turbidity that was measured at the time that the \(DF\) sample was taken. This was done for each moraine separately. The slope of the relation between the concentrations and the turbidity ranged between 2.9 and 3.3 mg l\(^{-1}\) NTU\(^{-1}\), which is comparable with values reported for Alpine streams (Geilhausen et al., 2013; Orwin et al., 2010; Paschmann et al., 2017). The coefficient of determination \((R^2)\) was 0.79 or higher. This correlation was used to convert the turbidity data to sediment concentrations in order to obtain high temporal resolution estimates of the suspended sediment concentrations. The suspended sediment flux was calculated by multiplying these estimated suspended sediment concentrations by the \(DF\) rate. For the 30-ky moraine at Sustenpass the sediment flux might be higher than estimated this way because some of the transported sediment got trapped in the runoff system and was thus not sampled. The total suspended sediment yield for each plot is defined as the sum of the sediment yield for the three sprinkling experiments. For the calculation of erosion rates, we assume that the sediment yield is equal to plot erosion.

### 3.3.4. Statistical Analyses

We determined the normality of the \(K_{\text{sat}}\) and soil aggregate stability data using the Shapiro-Wilk-test. We used the Kruskal-Wallis tests and Nemenyi tests to determine if the differences in the median \(K_{\text{sat}}\) and ASC values for the different moraines were statistically significant. We similarly tested the normality of the \(DF\) characteristics and whether the median values were statistically significant different for the three rainfall intensity classes (LI, MI, and HI). We used the Mann-Whitney test to evaluate if differences in the median values for the \(DF\) characteristics for each moraine age class were significant between the two study areas. To determine the correlations between surface characteristics and the \(DF\) characteristics, we used the Spearman rank correlation coefficient. We used a
0.05 level of significance for all analyses. All analyses were completed using the software R (v3.5.1—used with RStudio v1.1.463) and in particular, the packages: “stats” and “ggplot2.”

4. Results

4.1. Soil Surface Characteristics

The median vertical saturated hydraulic conductivity ($K_{sat}$) at the soil surface increased significantly from the oldest moraines (median of 540 mm hr$^{-1}$ on Sustenpass and 900 mm hr$^{-1}$ on Klausenpass) to the youngest moraines (median values of 4,320 mm hr$^{-1}$ on Sustenpass and 6,140 mm hr$^{-1}$ on Klausenpass). The median surface $K_{sat}$ values were generally higher at the Klausenpass than the Sustenpass but none of the differences between the similarly aged moraines was statistically significant (Figures 5a and 5b). A similar trend was seen for the $K_{sat}$ of the subsurface, but the median $K_{sat}$ values were an order of magnitude lower than at the surface. The median $K_{sat}$ at 5–20 cm below the soil surface was 8 and 20 mm hr$^{-1}$ for the 10-ky and 13.5-ky moraine versus 194 and 191 mm hr$^{-1}$ for the 30-y and 80-y moraine at the Sustenpass and Klausenpass, respectively (see Figures 4a–4d in Maier et al., 2021).

At Sustenpass, the median soil ASC (Figure 5c) decreased significantly from the 10-ky moraine to the 30-y moraine (median ASC of 0.94 and 0.22, respectively). At the Klausenpass (Figure 5d), the median ASC values were significantly higher for the two oldest moraines than the two youngest moraines (0.96 and 0.91 for the 13.5-ky and 4.9-ky moraine vs. 0.3 and 0.43 for the 160-y and 80-y moraine, respectively).

The average hydrophobicity on the soil surface was highest for the oldest moraines. The average $R$ was 5.25 for the 10-ky moraine at Sustenpass but only 1.22 for the 30-y moraine. At Klausenpass, the average $R$ was 9.36 for the 13.5-ky moraine and 1.72 for the 80-y moraine (Table 1). The differences between the differently aged moraines at the two study areas were statistically significant.

4.2. Overland Flow Response Characteristics for the Sprinkling Experiments

4.2.1. Runoff Ratios

For both study areas, OF was less frequently observed for the older moraines than the younger moraines (56%, 30%, 67%, and 78% of all experiments on the 10-ky, 3-ky, 160-y, and 30-y moraine at Sustenpass and 0%, 0%, 90%, and 67% of all experiments on the 13.5-ky, 4.9-ky, 160-y, and 80-y moraine at Klausenpass). The total OF volume for the 18 experiments per moraine age class ranged between 7 mm on the second oldest moraines and 40 mm on the second youngest moraines. The average runoff ratios were comparable for the different moraines (2%–5%), but peak runoff ratios increased with moraine age from 11% on the 30-y moraine to 42% on the 10-ky moraine at Sustenpass (Figures 6, 7, and 8j–l). The runoff ratio was not affected by the rainfall intensity when all experiments are considered together (Figures 8j–8l), but for the three oldest moraines at Sustenpass, the average runoff ratio decreased with sprinkling intensity (from 16% to 13% for the 10-ky moraine, from 6% to 4% for the 3-ky moraine, and from 5% to 3% for the 160-y moraine). This is reflected in the linear relation between the calculated effective infiltration rate and the applied rainfall intensity (Figure 9). There was no significant correlation between the runoff ratio and any of the measured surface characteristics (i.e., vegetation cover, vegetation complexity, $K_{sat}$, hydrophobicity, soil aggregate stability, rock cover, microtopography, and slope).

4.2.2. Peak Flow Rates and Drainage Volume

Peak flow rates ($Q_{peak}$) increased with rainfall intensity (average of 3.0, 6.4, and 8.2 mm hr$^{-1}$ for LI, MI, and HI sprinkling experiments, respectively) and with moraine age (from 4.6 mm hr$^{-1}$ for the youngest to 11.3 mm hr$^{-1}$ for the oldest moraines) at both study areas (Figures 6 and 8d–8f). The average volume of water that drained after rainfall stopped ($D$) did not depend on the rainfall intensity (Figures 8g–8i) but increased with moraine age; it was 1.8 mm for the oldest moraines and 0.4 mm for the youngest moraines.

4.2.3. Timing Characteristics

The average time until OF occurred ($t_{lag}$) decreased with rainfall intensity from 56 min for the LI experiments to 20 min for HI experiments (Figures 8a–8c) but did not clearly depend on moraine age and was similar for the two study areas. The average amount of rainfall applied before OF occurred ($P_{lag}$) ranged from 17 mm (oldest moraines) to 24 mm (second oldest moraines) and was not affected by rainfall intensity (Table S1).
4.3. Hydrograph Separation

The scaled EC values did not change notably during the experiments on the oldest moraines at Sustenpass (Figure 6b). The average scaled EC was stable at 0.5 (range: 0.30–0.75) for the 10-ky moraine and at 0.43 (range: 0.20–0.65) for the 3-ky moraine, suggesting a stable event water fraction of ∼50% (10-ky) and ∼43% (3-ky). This compares reasonably well with the isotope hydrograph separation results during the HI experiments (average of 41 ± 6% on the 10-ky and 44 ± 5% on the 3-ky moraines, respectively). The scaled EC changed during the experiments on the youngest moraines at both study areas (Figures 6b and 6d), which is reflected in the higher event water fractions ($f_e$) for the young moraines (60%–93%; Figure 8o).

At Sustenpass, the average event water contribution to OF decreased with moraine age from 79 ± 7% on the youngest moraine to 50 ± 3% on the 10-ky moraine. The average event water fractions for the two youngest moraines at Klausenpass were similar to the 30-y moraine at Sustenpass (77 ± 3%). The average event water

Figure 5. Boxplots of the saturated hydraulic conductivity ($K_{sat}$) of the surface (a and b) and soil aggregate stability (c and d) for the different moraines at Sustenpass (a and c) and Klausenpass (b and d). The box represents the 25th to 75th percentiles, the solid thick line the median, and the dashed line the average. The whiskers extend to the 10th and 90th percentiles. The symbols (jittered for better visualization) represent the actual measurements for the low, medium, and high complexity plots. Different capital letters above the box-plots denote statistically significant different median values for the different moraines in each study area. Note the log-scale for the $K_{sat}$ data (a and b).
Figure 6. Time series of the scaled overland flow (OF) rate and the scaled Electrical Conductivity (EC) of OF for the high intensity (HI) sprinkling experiments for all plots at Sustenpass (a and b) and Klausenpass (c and d) for which OF was observed. OF rates are scaled by dividing the flow rate by the average rainfall intensity for the experiment; EC values are scaled by dividing the measured EC by the EC of the applied rainfall. The colors represent the different moraine ages; the line type reflects the different plots (low, medium, and high vegetation complexity) on each moraine. Missing lines indicate that OF was not observed. Note that for the low complexity plot on the 30-y moraine at Sustenpass, data are only available for the first 60 min due to a broken hose that flooded the plot (as indicated with the // symbol). For the time series of the low and medium intensity (LI and MI) experiments, see Figure S5 and Figure S6.

Figure 7. Bar charts of the average overland flow (OF) runoff ratios (%) for the sprinkling experiments at Sustenpass (top) and Klausenpass (bottom) for each rainfall intensity class (LI, MI, and HI). The error bars indicate the minimum and maximum runoff ratios per intensity and moraine. Note that the y-axis for Sustenpass has a break between 26% and 38%.
Figure 8.
fraction including both study areas was significantly negatively correlated with vegetation cover ($r_s = -0.58; p < 0.001$) and soil aggregate stability ($r_s = -0.46; p = 0.005$), and significantly positively correlated to rock cover ($r_s = 0.47; p = 0.004$) and $K_{sat}$ ($r_s = 0.58; p < 0.001$).

For the oldest moraine at Sustenpass, the contribution of event water decreased from the MI to the HI experiments (from $\sim 63 \pm 2\%$ to $\sim 41 \pm 6\%$), but on the youngest moraine it increased (from $\sim 75 \pm 5\%$ to $\sim 81 \pm 3\%$). When comparing the hydrograph separation results for all experiments at both study areas together, the average event water fractions decreased slightly from the LI (72 $\pm$ 10$\%$) to the HI (67 $\pm$ 2$\%$) experiments (Figures 8m–8o). For the MI experiments and all of the HI experiments (except for the 30-y moraine with medium vegetation complexity), the soil water and previous event water fractions were notable and larger at the beginning of the event than at the end of the event (Figure 10).

4.4. Turbidity and Suspended Sediment Yield

The turbidity and suspended sediment concentrations were low and not notably affected by the rainfall rate for the three oldest moraines at Sustenpass (Figure 11). Peak turbidity and total sediment yield were much higher for the youngest (i.e., 30-y) moraine at Sustenpass. Peak turbidity for this moraine increased with rainfall intensity from $\sim 52$ NTU (LI) to $>75$ NTU (MI) and $>100$ NTU (HI). There was a positive correlation between rainfall intensity and average turbidity as well. This positive correlation between rainfall intensity and average turbidity was also observed for the youngest and second youngest moraine at Klausenpass. Peak turbidity was higher at Klausenpass than at Sustenpass and peaked at $>1,500$ NTU during the MI sprinkling experiment on the 160-y moraine at Klausenpass (Figure S12).

Figure 8. Overland flow (OF) response characteristics for all low (LI), medium (MI), and high (HI) intensity sprinkling experiments (left, middle, and right column, respectively) for each moraine age class (Sustenpass: circles; Klausenpass: triangles). The dashed line represents the average value for each sprinkling intensity class. Significant differences in the average values for the different rainfall intensities are indicated with a black star, significant differences between the Sustenpass and Klausenpass study areas are indicated by a gray star, and significant differences between the age classes by a brown star. Note that only the results for the experiments that resulted in OF are shown. $T_{OF}$ (a–c) is the time until OF occurred, $Q_{OF}$ (d–f) is the peak OF rate, D (g–i) is the total OF drainage volume, the runoff ratio (j–l) is the amount of rainfall that contributed to OF and $f_e$ (m–o) is the event water fraction in OF. Figure S7 shows the same results organized by moraine age class.
Figure 10.
For almost all of the experiments, the turbidity increased during the rising limb and remained stable during the recession, resulting in anti-clockwise hysteresis in the relation between the OF-rate and turbidity (Figure 11). The values of the hysteresis index $H$ (Zuecco et al., 2016) varied between $-0.69$ and $0.37$ for the LI experiments, between $-0.65$ and $0.10$ for the MI experiments and between $-0.60$ and $0.16$ for the HI experiments (average of $-0.29$ for all the experiments).

The average suspended sediment yield for the youngest moraines increased with rainfall intensity from 3 g (1–5 g) to 9 g (1–28 g) and 14 g (2–46 g) for the 30-y and 160-y moraine at Sustenpass and from 17 g (4–24 g) to 113 g (20–530 g) and 164 g (3–796 g) for the 80-y and 160-y moraine at Klausenpass. The total suspended sediment yield for each experiment was significantly correlated with the OF volume ($r_s = 0.87$; $p < 0.001$ when analyzing the data for both study areas together). There was no correlation between the slope of the plots and total sediment yield per plot, but there was a significant negative relation between vegetation cover and total sediment yield.

Figure 10. Overland flow (OF) hydrographs and hydrograph separation results for the high intensity (HI) sprinkling experiments at Sustenpass (a) and Klausenpass (b). The hydrographs for the different plots (i.e., subpanels) are arranged by moraine age (rows; old [top] to young [bottom]) and vegetation complexity (columns; low complexity [left] to high complexity [right]). The contributions of the different components are visualized with different colors (orange = soil water before any of the sprinkling experiments; dark blue = rainfall applied during the previous [i.e., MI] experiment; light blue = rainfall applied during the experiment). The arrows in the panels mark the end of the sprinkling. Note. That the scales of the $x$- and $y$-axes differ for each panel to better visualize the differences in the hydrograph separation results. See Figure S8 for the results with a similar $x$-axis scaling and Figures S9–S10 for the hydrograph separation results for the LI and MI experiments. Figure S4 shows the mixing diagrams.

Figure 11. Time series of the turbidity of overland flow (OF) for the high intensity (HI) sprinkling experiments at Sustenpass (top) and Klausenpass (bottom). The colors represent the different moraines and the line style the different plots (low, medium, and high vegetation complexity). The size of the circles and the number inside represent the total sediment yield for each experiment (in g). The outlines of the circles represent the plot (similar as for the time series). For the results for the low and medium intensity (LI and MI) experiments, see Figures S11 and S12, respectively.
The total sediment yield was also negatively correlated with the soil aggregate stability ($r_s = -0.35; p = 0.009$).

### 4.5. OF During Natural Rainfall Events at Klausenpass

OF occurred mainly during periods of high rainfall intensity during long rainfall events (e.g., the 95 mm event between the 11th and 13th of August; Figure 12). Shorter events with a higher intensity did not cause a strong OF response (e.g., the 58 mm event on the 7th of August). During the first major natural rainfall event, the OF response to rainfall was quick for all of the moraines (Figure S13). However, during the second and the third event, there was a time lag of several hours between the onset of the rainfall and the OF response, but OF was generated rapidly after subsequent bursts of rainfall.

The average OF ratios for the four largest events (Jul 27th–29th; Aug 6th–8th; Aug 11th–13th; Aug 19th–21st) during the measurement period decreased with moraine age (5%, 6%, 4%, and 2% for the 80-y, 160-y, 4.9-ky and 13.5-ky moraine, respectively). The OF ratios were highest (8%) for the 160-y and 80-y moraine plots for the events on Aug 19th–21st and Jul 27th–29th, respectively. These OF ratios are comparable with the OF ratios of the LI and MI sprinkling experiments (4%–8%; Figure 7). No OF was observed during the sprinkling experiments at the 13.5-ky, 4.9-ky, and 80-y low complexity plots and thus a comparison between the runoff ratios during natural events and sprinkling events cannot be made for these plots.

OF was only generated when a threshold for the volume of rainfall plus the antecedent soil moisture storage (ASI) in the top 10 cm of the soil was exceeded (Figure 13). For the low complexity plots on the 13.5-ky, 4.9-ky, and 80-y moraine, these thresholds were 80, 77, and 47 mm, respectively. Almost every event led to an OF response for the 160-y moraine, and a storage threshold was thus not observed. The higher rainfall thresholds for the older moraines are in line with the higher volumetric water content at 10 cm soil depth for the 13.5-ky (average: 0.26; range: 0.22–0.31) and 4.9-ky (average: 0.29; range: 0.25–0.33) moraines compared to the 160-y moraine (average 0.08; range: 0.06–0.10) and 80-y moraine (average 0.09; range: 0.08–0.11) during the July 25, 2019–September 15, 2019 study period. At Sustenpass, the mean volumetric water content during the field season was also higher for the oldest moraine (average: 0.38; range: 0.35–0.41) than the other moraines (average values of 0.15, 0.12, and 0.10 for the 3-ky, 160-y, and 30-y moraine, respectively).
5. Discussion

5.1. Changes in Surface Characteristics With Moraine Age

As expected, plant cover increased with moraine age at both study areas, except for the 3-ky moraine at Sustenpass, where rockfall resulted in fields of scree and reduced the vegetation cover for two plots (Figure 3 and Table 1). The associated increases in root density, macroporosity and preferential flow (Hartmann, Semenova, Weiler, & Blume, 2020; Maier et al., 2020) are normally expected to cause an increase in the saturated hydraulic conductivity \( K_{sat} \) (Beven & Germann, 1982; Weiler, 2017; cf. arrow #6 in Figure 1). However, the vertical \( K_{sat} \) decreased with moraine age at both the Sustenpass and Klausenpass (Figures 5a and 5b). This is probably caused by the changes in soil texture (see #1 in Figure 1), particularly the increase in silt and clay content (#4 in Figure 1) and decrease in gravel content with moraine age (Hartmann, Weiler, & Blume, 2020; Maier et al., 2020; Musso et al., 2019) due to physical and chemical weathering (D’Amico et al., 2014; Egli et al., 2001; Mavris et al., 2010). The comminution of soil particles by congelifraction can result in rapid changes in soil texture in Alpine areas (Oztas & Fayetorbay, 2003). Such an increase in fine particles during soil development has been reported in other studies (e.g., D’Amico et al., 2014; Egli et al., 2001; Mavris et al., 2010; Roberts et al., 1988). It also has a large effect on the soil water retention capacity (Hartmann, Semenova, Weiler, & Blume, 2020) and the average soil moisture content.

One would expect the lower \( K_{sat} \) to lead to less infiltration and more OF (#10 in Figure 1) but the calculated effective infiltration rates (Figure 9) were similar for the different moraines and depended mainly on the rainfall intensity. There is, however, a hint that the effective infiltration rates for the oldest moraines are slightly lower than for the younger moraines (i.e., the points plot below the line in Figure 9), but this effect is small and remains suggestive.

For the oldest moraines at Sustenpass and Klausenpass, the \( R \) values were higher than 1.95, which suggests a negative effect on the sorptivity (Tillmann et al., 1989). The denser cover of Alpine rose for the oldest moraines (as well as the 4.9-ky moraine at Klausenpass) and related release of hydrophobic root exudates (Eilnenberg & Leuschner, 2010), may have increased the hydrophobicity and reduced infiltration rates (#5 in Figure 1). These results are in line with findings of previous studies that report higher \( R \) values for forests and shrubland than

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Figure 13. Event total overland flow (OF) \( (Q_{total}, \text{mm}) \) for each sprinkling experiment (triangles) and natural rainfall event (diamonds) as a function of the sum of the rainfall and the average soil water storage in the top 10 cm of the soil (ASI, mm) before the start of the event. The different colors represent the different low vegetation complexity plots at the Klausenpass. Note. For the 4.9-ky and 80-y moraine plot, there is no soil moisture data available for the natural rainfall event on Jul 27th–29th.
grasslands and meadows (Badoux et al., 2006; Lichner et al., 2007). One would expect that this leads to more OF (#5 and #10 in Figure 1) but this was not the case as $Q_{\text{total}}$ (Table S1) and the average runoff ratios (Figure 7) were not higher for the oldest moraines.

Chemical weathering is accelerated by vegetation growth and the release of root exudates that break down primary soil minerals and stimulate the formation of secondary minerals (#1 in Figure 1). The binding forces of the clay minerals likely stabilized the soil particles on the oldest moraines (Lado et al., 2004; cf. #2 in Figure 1), whereas on the youngest moraines, coarse gravel fragments cause poor aggregation. In addition to the chemical weathering, the higher soil organic matter and root density on the older moraines (Maier et al., 2020) likely also resulted in more stable soil aggregates (Figures 5c and 5d; cf. #3 in Figure 1), and increased the physical enmeshment of particles and root mucilage (Morel et al., 1991). This likely reduced slaking and the wettability of the aggregates (Le Bissonnais et al., 2018; Sullivan, 1990), leading to the lower suspended sediment concentrations for the oldest moraines (cf. #7 in Figure 1).

The more gradual change in $K_{\text{sat}}$ and soil aggregate stability with moraine age at Sustenpass compared to Klausenpass is likely due to the lower pH of the moraines derived from silicate bedrock that lead to stronger chemical reactions and a faster development of secondary clay minerals (Pott & Hüppe, 2008). This also explains why the soils on the oldest moraines at Sustenpass were finer and more well-developed (see also Maier et al., 2021), and the overall higher soil moisture content and water retention for the oldest moraines at Sustenpass compared to Klausenpass (Hartmann, Semenova, Weiler, & Blume, 2020; Hartmann, Weiler, & Blume, 2020).

### 5.2. Overland Flow Generation

#### 5.2.1. Overland Flow on Old Moraines

Surface $K_{\text{sat}}$ values were lower for the old moraines than the young moraines (Figures 5a and 5b), but still larger than the intensity of a 100-yr 1-h rainfall event (∼50 mm hr$^{-1}$). This suggests that OF at the old moraines was very unlikely to be HOF. However, clay-rich layers were observed at 20–40 cm below the surface of the old moraines in both study areas (Hartmann, Weiler, & Blume, 2020; Maier et al., 2021), which may have limited infiltration and caused SOF (cf. Holden & Burt, 2003; Mayerhofer et al., 2017). The development of the pronounced organic surface layer and dense root network will have allowed for rapid lateral drainage of these surface layers, so that biomat flow may have been important as well (cf. #9 in Figure 1). Brilliant blue dye staining experiments by Hartmann, Semenova, Weiler, and Blume (2020) and Hartmann, Weiler, and Blume (2020), indeed, showed that the infiltration depth on the oldest moraine at Sustenpass is limited and infiltration is mainly driven by macropore flow via root channels (cf. #9 in Figure 1). The large volume of water that drained from the slope after the end of the sprinkling experiments ($D$) for this moraine (Figure S7) also hints at the importance of lateral drainage of the organic rich surface layers. Biomat flow has previously been observed for pre-Alpine grasslands (Scherrer et al., 2007).

For the oldest moraines at Sustenpass, mean OF ratios were higher for the MI sprinkling experiments than the HI experiments (Figure 7). This is at first sight perhaps counter-intuitive as higher rainfall intensities usually produce more OF. Instead, the calculated effective infiltration rates were higher for the higher intensity sprinkling experiments (Figure 9). This is most likely due to the spatial variability in infiltration rates caused by macropore flow (Hartmann, Semenova, Weiler, & Blume, 2020; Maier et al., 2020). In addition, Alpine rose on the 10- and 3-ky moraine may possibly have provoked hydrophobic topsoil conditions for the first (i.e., LI) experiments (Table 1; Ellenberg & Leuschner, 2010). Potentially, these water-repellent films were removed by the sprinkling experiments, leading to enhanced infiltration and lower OF ratios for the later higher intensity experiments.

The higher moisture storage and smaller drainable porosity for the older moraines at Sustenpass led to much more mixing of the rainfall with the stored soil water than for the younger moraines (Figures 8m–8o and 10). The pre-event water fractions in OF increased over time, which suggests that flow through old root channels or biomat flow were particularly important for the first sprinkling experiments and early on during the events but that exfiltration of soil water was more important later during the event (i.e., when the soil had become field-saturated and any water-repellent coatings had been washed off). Large pre-event water fractions in OF have been reported in other studies as well (e.g., Ribolzi et al., 2000). Sprinkling experiments on meadows overlying Cambisols in the Swiss pre-Alpine area indicated that OF consisted of ∼7% to ∼26% pre-event (soil) water but that this percentage decreased during the experiments (Kienzler & Naef, 2008).
In contrast to the old moraines at Sustenpass, sprinkling experiments on the old moraines at Klausenpass did not produce any OF. Presumably this is related to the less developed soils on the old moraines at Klausenpass compared to those at Sustenpass. The soils of the old moraines at Klausenpass had a lower organic carbon content (Hartmann, Weiler, & Blume, 2020) and less-pronounced clay-rich layers compared to the old moraines at Sustenpass. This led to a generally lower soil moisture content and higher drainable porosity, and a higher $K_{\text{sat}}$ at depth (see also Figure 4 in Maier et al., 2021). This will have resulted in a higher rainfall plus ASI threshold for OF, and thus less frequent ponding above this layer, and consequently less frequent OF. In other words, the sprinkling experiments were not big enough to saturate the topsoil and cause OF. During natural rainfall, the intensity (1–3 mm hr$^{-1}$ for the different events) was much lower than for the sprinkling experiments but the rainfall volume was much larger. The large rainfall volume caused saturation of the uppermost soil layers and OF once the storage threshold was exceeded (Figure 13; cf. arrow #11 in Figure 1). Topsoil saturation has been observed by previous studies as well (e.g., Brown et al., 1999; McDonnell et al., 1991). For example, Kienzler and Naef (2008) reported limited infiltration into the subsurface despite high macroporosity of the soil and showed that the uppermost part of the soil became saturated nearly independently from the subsoil. This saturation was caused by a thick organic horizon. The very dense layer of roots in the top ~5 cm of the soil on the oldest moraines at Klausenpass could have become saturated and promoted lateral flow. This saturation was probably too localized and too close to the surface to be detected by the soil moisture sensors at 10 cm depth. The delay between the start of the rainfall and the onset of OF (Figure S13) represents the time that is needed to saturate the top soil and to reach the soil moisture threshold that is needed to trigger SOF during a burst of high intensity rainfall (i.e., that is more intense than the deeper percolation rate).

### 5.2.2. Overland Flow on Young Moraines

Even though the measured surface $K_{\text{sat}}$ values for the young moraines were very high (Figures 5a and 5b) and much larger than the rainfall intensities, OF occurred for many natural rainfall events (only measured at Klausenpass) and most of the sprinkling experiments. The high $K_{\text{sat}}$ values suggest that OF is not due to Hortonian OF (HOF). The high $K_{\text{sat}}$ values at 5–20 and 20–40 cm below the soil surface and the low average soil moisture content in the top 10 cm of the soil suggest that saturation-excess OF (SOF) is also unlikely.

On some plots (e.g., the low complexity plot on the 160-y moraine at Klausenpass), surface sealing and crusting of the loose sediment was observed after the first sprinkling experiment. This may have led to lower infiltration rates and higher OF in the subsequent experiments than the $K_{\text{sat}}$ values would suggest. This phenomenon has been reported in various other studies (Badoreck et al., 2013; Huang et al., 2002; Onda et al., 2008). For example, Schaal et al. (2013) observed lower infiltration rates and higher OF during early stages of landscape development due to crusting of unvegetated coarse-grained material.

The high stone- and rock cover of the young moraines (Figure S14) are likely important for OF generation (cf. Descroix et al., 2001; Mayor et al., 2009; Poesen et al., 1990; Valentin, 1994; #8 in Figure 1). Especially large, closely spaced stones and rocks (Lavee & Poesen, 1991) near the bottom of the plots may act as an impermeable layer, from which water can flow to the collection gutters (Figures 3 and S14). This may have caused the OF on the 30-y moraine at Sustenpass and the 160-y and 80-y moraine at Klausenpass. The maximum OF runoff ratio on the young moraines was 11% (minimum: 1%), which would suggest that indeed only a small proportion of the plot (max 2.6 m$^2$) contributed to OF generation. OF generation from the stones and rocks near the plot bottom would explain the quick onset of OF for some experiments on the young moraines (min $t_{\text{lag}}$ of 5 min on the 160-y moraine at Klausenpass and 9 min on the 80-y moraine) as well. However, this does not fit with the delay of several hours between the start of the rainfall and the onset of OF for the natural rainfall events with low rainfall intensity (Figure S13). These long lag times, especially for very long and large rainfall events have been observed for other sprinkling experiments as well (Gomi, Sidle, Ueno, et al., 2008; Yu, 1999). The substantial drainage after the end of the sprinkling experiments also doesn’t fit with OF that is caused by rainfall falling on the rocks near the gutters. Instead, the stones and rocks that are buried near the surface may have caused local ponding and saturation of the top soil and an OF response after a certain storage threshold was exceeded (Figure 13). Most likely, saturation of the soil between the very large rocks during the bursts of higher intensity rainfall (Figure S13) was limited to the top few centimeters of the soil. The decrease in OF rates during some sprinkling experiments (Figure 10) suggests that infiltration in the finer material between the rocks became faster during the event, either because preferential flow pathways became connected, or the effect of surface sealing decreased.
Due to the low water holding capacity of the coarse and poorly aggregated sediment at the young moraines and thus limited pre-event water storage, there was little mixing of the rainfall with soil water. The large event water contributions to OF (Figures 8m–8o and 10) and increases in the EC during the HI experiments (Figures 6b and 6d) also suggest that OF was largely generated by water flowing on the surface and/or over rocks very close to the surface. Although a comparison to other studies of OF on very young soils in proglacial environments is not possible, the high fractions of event water in OF on the young moraines seem reasonable. Cras et al. (2007), for example, analyzed flash-flood events on porous, unvegetated marl soils in badland areas in the Southern French Alps and found that surface flow pathways were the main mechanism of runoff generation and resulted in high event water contributions for streamflow. In Australia, between 65% and 75% of the runoff in a degraded, semi-arid catchment with shallow soils was derived from rainfall (Turner et al., 1991).

5.2.3. Effects of Moraine Age on Overland Flow Characteristics

The results of this study show that soil and vegetation development affect hillslope characteristics and change OF generation (cf. Figure 1; for a visual representation of these processes see also Figure 12 in Maier et al., 2021). Average OF ratios for all sprinkling experiments were similar (2%–5%) because on the young moraines OF occurred frequently but the runoff ratio was low, while on the old moraines OF was less common but runoff ratios were high when it occurred. On almost all moraines at Klausenpass, OF occurred only after a rainfall plus soil water storage threshold had been exceeded (Figure 13).

However, the OF generation mechanisms were different. On the young moraines, OF is mainly produced by SOF and rainfall falling on rocks and stones near the surface. On the older moraines the development of organic matter, root mats and a clay rich layer are decisive for SOF generation (cf. Lohse & Dietrich, 2005). While OF is rapid for the young, freely draining moraines and dominated by rainfall, soil water exfiltration is important for OF on the old moraines. The better moisture retention for the old moraines (i.e., higher moisture content at field capacity; Hartmann, Weiler, & Blume, 2020) results in a larger volume of water stored in the soil that can mix with the rainfall (cf. Lohse & Matson, 2005). It is also evident that, especially after dry periods, the impact of hydrophobicity should not be underestimated but the type of water repellency changes from sand hydrophobicity and crusting on the young moraines to hydrophobicity driven by plant root exudates on the old moraines.

Although the changes during landscape evolution were similar for the two study areas, the type of bedrock controls the rate of the changes in soil and vegetation characteristics and consequently the modification of OF characteristics over time (cf. #11 in Figure 1). The quicker and more pronounced soil development at Sustenpass resulted in more and more frequent OF for the old moraines at Sustenpass than at Klausenpass.

5.3. Changes in Sediment Yield With Hillslope Age

The turbidity measurements and sediment yield calculations (Figure 11) suggest that OF driven soil erosion is only considerable for the young, mainly unvegetated moraines (namely 30-y moraine at Sustenpass and the 160-y and 80-y moraine at Klausenpass). The variation in the slope of the plots did not explain the variation in the OF volumes, nor the sediment yield. The OF volume, furthermore, did not explain the changes in sediment transport with moraine age because the OF volumes did not differ significantly between the moraine age classes (Figure S7). Rather the effects of vegetation cover and soil aggregate stability explain the variation in sediment yield along the chronosequences (cf. #3 and #7 in Figure 1). A negative correlation between sediment yield and the vegetation cover has been observed in several other studies (e.g., Abrahams et al., 1995; Emmett, 1970). Vegetation increases the critical shear stress for sediment transport (Prosser et al., 1995), thereby reducing the sediment yield. The increases in organic matter and clay content for the older moraines (Greinwald et al., 2021; Hartmann, Weiler, & Blume, 2020), furthermore, increase soil aggregate stability (see #2 and #3 in Figure 1 and Figure 5), which reduces the susceptibility for erosion (cf. Barthès & Roose, 2002; Le Bissonnais et al., 2002). The observed decrease in sediment yield and stronger dependence on soil aggregate stability and vegetation cover than rainfall intensity can be implemented in landscape evolution models using the concepts of transport- and detachment limitation (Carson & Kirkby, 1972; Howard, 1994; Whipple & Tucker, 2002). More specifically, it suggests that the increase in the detachment threshold over time is more pronounced than the increase in the diffusion coefficient (which is a function of the surface water flux; Temme et al., 2017).

Despite the low vegetation cover and soil aggregate stability, sediment transport on the youngest moraines may be limited by the sediment supply. Most of the plots on the youngest moraines and the 160-y moraine at Klausenpass...
have more than 20% stone and rock cover (Figure 3), above which sediment flux tend to decrease (Bunte & Poesen, 1993). Most of the stones and rocks on the young moraines are located on top of the surface and not buried in the soil. This generally results in less OF and less sediment transport, even during very intense rainfall events (Li et al., 2020; Mandal et al., 2005). The anti-clockwise hysteretic relation between turbidity and OF for almost all of the experiments also suggests that erosion is transport limited and that it takes longer for sediment sources further away to be hydrologically connected to the gutter (Bača, 2008; Benkhaled & Remini, 2003; Klein, 1984). However, it might also indicate plunges of stones or even local collapses of microtopographic relief during the experiment and transport of larger amounts of sediment toward the bottom of the moraines (Russell et al., 2001; Sarma, 1986).

Musso et al. (2020) report long-term (∼60 yr) erosion rates of 4–5.5 t ha⁻¹ yr⁻¹ for the young moraines and 1–2.6 t ha⁻¹ yr⁻¹ for the old moraines at both study areas. These differences between the young and old moraines are in agreement with the large differences between the very high erosion rates on the young moraines and much lower erosion rates on the old moraines for both study areas. A review of erosion rates in the Alps by Meusburger (2010) suggests typical soil loss rates between 2 and 33 t ha⁻¹ yr⁻¹, even for grassland sites. Descroix and Mathys (2003) reported for degraded land with similar slope characteristics as the young moraines sediment fluxes of 4–33 t ha⁻¹ yr⁻¹. If we assume that there is one event per year with the same sediment flux as the experiment with the largest sediment flux (HI event on the low complexity plot of the 160-y moraine at Klausenpass) and assume that sediment delivery equals soil erosion, then the estimated erosion rate would be ∼330 kg ha⁻¹ yr⁻¹, which is of the same order of magnitude as that reported by Musso et al. (2020). Admittedly, this event would not occur each year, and there would be other events that lead to erosion as well. Furthermore, there is significant deposition of the sediment within the plots and the erosion rates are thus larger than the measured sediment yield. Lustenberger (2019) used colored sand to study sediment movement on the plots at Sustenpass and showed that there was indeed significant redistribution of the sediment within the plot and that the sediment travel distances were larger for the 30-y moraine than the 160-y moraine and 3-ky moraine.

6. Conclusions

Our experimental study on moraine chronosequences in two proglacial areas in the Central Swiss Alps shows that despite the increase in vegetation cover, root density, and soil organic matter content, the saturated hydraulic conductivity decreased with moraine age due to the decrease in gravel and increase in silt and clay content. However, this did not affect the effective infiltration rates during the sprinkling experiments, which was high for all moraines and increased with rainfall intensity.

The peak OF ratios increased with moraine age and were largest for the oldest moraines on silicate bedrock, where OF was driven by flow along the dense root network and saturation-excess due to ponding above a lower permeability clay-rich layer at 20–40 cm below the soil surface. In the calcareous glacier forefield, however, OF did not occur on the old moraines during the intense sprinkling experiments, presumably because the clay-rich layers were less pronounced and the sprinkling experiments were too small to cause saturation. During larger (but lower intensity) natural rainfall events that exceeded the storage threshold, OF was observed for these moraines. OF occurred most frequently on the young moraines and was likely caused by saturation of the soil above stones and rocks that were buried near the surface, even though the infiltration rate in the sediment between the stones was very high. The larger water storage in the soils of the older moraines, due to the higher water retention caused by the larger silt-, clay-, and organic matter content, resulted in more mixing of rainfall and soil water and a larger contribution of pre-event water to OF than for the younger moraines. The higher soil aggregate stability and vegetation cover—rather than differences in the OF volume or flow rates—caused the sediment fluxes to be lower for the old moraines than the young moraines.

These results, in combination with the changes in SSF (Maier et al., 2021), help us to understand how hillslope characteristics and runoff generation processes change during the first millennia of soil development and the interactions between these processes (cf. Figure 1). This is important for understanding the spatial and temporal changes in runoff generation in rapidly changing Alpine areas and can be used to inform landscape evolution models.
Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Data Availability Statement

The data described in this article is openly available in the online data repository GFZ Data Services: https://dataservices.gfz-potsdam.de/pnmetaworks review/51c19bb26c936d447a5263dbb091d9b525a8f16b39efb-7306413e797eeea488/8

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