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1 **Vegetation influence on soil formation rate in a proglacial chronosequence (Lys Glacier, NW**
2 **Italian Alps)**

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7

8 **Abstract**

9 Climate change has huge impacts on alpine ecosystems . One of the most visible effects in the Alps
10 is glacier retreat since the end of the Little Ice Age (LIA), which caused the exposure of previously
11 glaciated surfaces. These surfaces are open-air laboratories, verifying theories regarding ecosystem
12 and soil development.

13 In order to increase our knowledge on the effects of time and vegetation primary succession on soil
14 development in proglacial areas, we sampled soils and surveyed plant communities on stable points
15 in the proglacial area of the Lys glacier, in the Italian north-western Alps (Valle d'Aosta Region).

16 The sampling points were located on dated sites (based on literature and/or historical photographs).

17 Glacial till is attacked by weathering processes immediately after deposition and stabilization, with
18 a consequent loss of soluble compounds, decrease of pH and primary mineral weathering. The
19 speed of these processes was largely increased after the establishment of a continuous vegetation
20 cover, thanks to the organic matter accumulation caused by litter input and root decomposition
21 below the soil surface.

22 On silic glacial tills, below timberline and under a quasi-climax Larch - Rhododendron forest, a
23 fast and steady decrease in pH values, increase in organic matter content and horizon differentiation
24 was observed. In particular, genetic eluvial horizons formed in approximately 60 years, while
25 diagnostic albic horizons were developed after ca. 90 years, pointing to an early start of the
26 podzolization processes. Cheluviation of Fe and, secondarily, Al were analytically verified.

27 However, illuviation of Fe, Al and organic matter in incipient B horizons were not sufficient to
28 obtain diagnostic spodic horizons on LIA materials.

29 Under grazed grassland below timberline and alpine prairie above timberline, acidification and
30 weathering were slightly slower, and no redistribution with depth of Fe and Al oxi-hydroxides was
31 observed. A cambic Bw horizon developed on the oldest LIA moraines.

32 Therefore it seemed that this fast onset of the podzolization process in comparison to other
33 proglacial chronosequences in the Alps was mainly driven by vegetation properties rather than by
34 specific climatic conditions.

35

36 **Keywords**

37 Soil formation; proglacial area; soil chronosequence; podzolization; subalpine soils; alpine soils

38

39 **Highlights**

- 40 • Plant succession and soil development were slower above than below timberline;
- 41 • Under grassland below and above timberline, Cambisols were found on old moraines;
- 42 • Under typical subalpine forest-shrub vegetation, podzolization was very fast;
- 43 • E horizons and “real” Podzols appeared faster than in other forefields in the Alps;

- Vegetation seemed very important in the early podzolization onset.

45

46 **1. Introduction**

47 Climate change is having huge impacts on mountain ecosystems. One of the most visible effect is
48 the glacier retreat, which continued with few interruptions since the end of the Little Ice Age (LIA,
49 between ca. years 1300 and 1821/1861, Ivy-Ochs et al., 2009). The bare surfaces left by retreating
50 glaciers offer the opportunity to observe the early stages of soil development and the primary
51 ecogenetic succession, thus validating existing theories about ecosystem evolution and allowing to
52 determine the speed of soil forming processes. A large number of soil chronosequences in proglacial
53 areas has been studied on mountain ranges worldwide and in arctic and antarctic environments (e.g.
54 Ugolini, 1966, Burt and Alexander, 1996, Egli et al., 2001, Dümig et al., 2011, Mahaney et al.,
55 2009, Hodkinson et al., 2003). According to most of these studies, soil chronosequences under
56 boreal/subalpine forests frequently end up with Podzols, but the rate of development of a podzolic
57 morphology and chemistry widely differ in several studies considering different environments. For
58 example, in superhumid basins in New Zealand, with an average annual rainfall higher than 10000
59 mm (Tonkin and Basher, 2001), Podzols were formed in about 1000 years, despite the extremely
60 fast weathering of the gravel and sand fractions and the consequent increase in silt and clay contents
61 in surface horizons. In Alaska, high rainfall and humidity leads to the formation of weakly
62 developed E and Bs horizons after only 70 years, and “real” Podzols after 230 years (Ugolini, 1966,
63 Burt and Alexander, 1996). In these areas, E horizons immediately appeared after the establishment
64 of spruce in the forest succession. This happened despite the presence of abundant carbonatic
65 materials in the parent till. Similar fast rates of Podzol formation (less than 230 years) were
66 observed on marine terraces in Finland (Mokma et al., 2004).

67 Focusing on the European Alps, many studies have been performed on soil and ecosystem
68 development in proglacial areas in subalpine forest habitats (e.g. Egli et al., 2003, Egli et al., 2006,
69 Dümig et al., 2011), while a smaller number was devoted to alpine (i.e. above timberline) habitats,
70 with most of them considering mainly primary plant succession (e.g., Andreis et al., 2001,
71 Caccianiga et al., 2006, Raffl et al., 2006).

72 In the fifties some works described Podzol-like soils on recently deglaciated surfaces also in the
73 Alps: for example, Jenny (1958) showed a 315 years old soil which had a 11 cm thick light gray
74 horizon overlying a 14 cm thick brownish one, under subalpine ericaceous shrubs in the Rhone
75 Gletscher forefield. However, in the alpine range, much longer periods since deglaciation were
76 normally required for the establishment of the podzolization process than in other boreal habitats.
77 Under subalpine vegetation, the youngest soils were usually Lithic and Skeletic Leptosols; after 120
78 years soil evolution led to the development of Humi-Skeletic Leptosols or, locally, to weakly
79 developed Dystric Cambisols (Righi et al., 1999, Egli et al., 2001, Egli et al., 2006). A continuous

80 humus layer appeared in the oldest soils, together with weak signs of chemical weathering
81 (formation of Fe and Al oxyhydroxides). Oxalate extractable Fe and Al decreased with depth, and
82 no cheluviation was visible (Egli et al., 2006). In the Alps, Dystric Cambisols are normally found
83 on 250-300 years old surfaces, while Podzols appear after 600 years in very humid sites (1800
84 mm/y, Zech and Wilcke, 1977), but usually more than 1300 years are needed for the development of
85 E horizons (Egli et al., 2003). In central Switzerland, in areas characterized by mean annual
86 precipitation equal to ~2000 mm/y, Podzols are found in 3300 years old sites (Egli et al., 2001,
87 Dümig et al., 2011), but they are normally weakly developed (Righi et al., 1999).

88 The establishment of vegetation on the bare deglaciated areas initiates gradients in many soil
89 properties. The primary succession of vegetation in several glacier forefields in the European Alps
90 include pioneer stages dominated by *Oxyrietum digynae* and *Epilobietum fleischeri* (Braun-
91 Blanquet 1948), lasting around 30 years, followed by the establishment of an open forest. *Larix*
92 *decidua*, *Rhododendron ferrugineum* and other Ericaceae start colonizing regolith surfaces after
93 about 20-30 years since deglaciation, but *Rhododendron* generally shows low cover values in most
94 Swiss chronosequences, where the LIA primary succession was normally dominated by *Alnus*
95 *viridis* (e.g., Burga, 1999, Föllmi et al., 2009, Burga et al., 2010) (except in the Damma glacier
96 forefield, Bernasconi et al., 2011).

97 Despite these well-established patterns of soil and plant community chronosequences, in many
98 glacier forefields of the Aosta Valley (North-western Italian Alps), both plant community succession
99 (particularly at the subalpine belt) and pedogenesis seem faster than in most other proglacial areas
100 in the Alps. For example, in the proglacial area considered in this study, *Rhododendron ferrugineum*
101 dominates the understory vegetation already after few decades from glacial retreat (Treter et al.,
102 2002). During many soil survey campaigns in Aosta Valley, a well-defined bleached E horizon was
103 commonly recognized in soils developed on 130 years old moraines, associated with weakly
104 developed BCs horizons, slightly enriched in pedogenic oxyhydroxides; this pattern was recorded
105 also in the Miage glacier forefield in the Mont Blanc Massif and in the Verra Grande glacier
106 forefield in the Monte Rosa Group. In these areas, above the treeline, ecosystem and soil
107 development had a similar rate as in other similar alpine environments (Andreis et al., 2001), and
108 “climax” soils and plant communities are reached in longer periods than at lower elevation.

109 Thanks to the well-known chronology of deglaciation of the Lys glacier (e.g., Monterin, 1932), we
110 chose its proglacial area to investigate the rates of the soil forming processes in the Alps. The Lys
111 forefield includes habitats both below (subalpine) and above timberline (alpine), thus evidencing
112 multidirectional trends in pedogenesis and vegetation succession (Huggett, 1998), hence
113 contributing to the detection of the main driving factors of these processes.

114 The comparison between contrasting subalpine and alpine primary vegetation successions and soil
115 chronosequences provides useful information to evaluate the importance of pedoclimatic and
116 vegetation spatial variability in driving the direction of pedogenic processes.
117

118 **2 Material and Methods**

119 **2.1 Study area** 120

121 The proglacial area of the Lys Glacier is located in the upper Lys Valley (Aosta Valley, Pennine
122 Alps, Italy, Figure 1). The morainic systems left by several glacier fluctuations during the Holocene
123 were usually erased by the larger advance of the Little Ice Age (LIA), between ca. years 1350 and
124 1850 (Joerin et al., 2006). The maximum LIA glacier advance was reached in 1821; a secondary
125 advance ended in 1861, when the glacier approached the moraine deposited in 1821, leaving no
126 frontal morainic arc because of river erosion. Since 1861, approximately 1.8 km were left free of
127 ice. A minor advance (1915-1921) left a small morainic arc about 800 m from present day glacier
128 terminus.

129 We sampled a recent soil (P1, ca. 7-10 years old) as representative of fresh, raw till (starting point of
130 soil development and vegetation succession for both alpine and subalpine chronosequences). Above
131 timberline (“alpine” chronosequence), we sampled soils formed at 4 different sites under common
132 vegetation covers (Table 1). Below timberline (subalpine chronosequence), we sampled soils at 5
133 different sites (Table 1); if two different vegetation types occurred on same-age surfaces, we opened
134 a soil pit under each type (i.e. S3, S4, S6 were observed under larch forest with *Rhododendron*
135 *ferrugineum*, SG2, SG5, SG7 below grazed grassland). S10 and A6 are on late-glacial (Younger
136 Dryas) till (more than 10000-11000 years old, according to Pelfini et al., 1997). This age is
137 attributed thanks to similarities with most other proglacial areas in the Alps (Ivy-Ochs et al., 2009).
138 We excluded areas visibly disturbed by erosion or deposition processes. All soil profiles were
139 chosen as representative among a much larger number of soil observations by soil coring. Since
140 small scale topographic differences have strong effects on soil development and vegetation
141 succession dynamics (Burga et al. 2010), sites with similar surface rockiness and slope were
142 chosen.
143 Unfortunately, below the present-day glacier front the slope is steep and eroded (“Rocce di
144 Salzen”), and it becomes reasonably stable only 200 m below, where the material was left by the
145 glacier around year 1945. Therefore, no soil has been sampled and analyzed with ages ranging
146 between 5/10 and 60/70 years.

147 The altitude of the proglacial area ranges from 1990 m a.s.l. to about 2480 m a.s.l.. The highest
148 morainic ridge (2480 m a.s.l.) was deposited in 1755 (Strada, 1988). The present-day glacier tongue
149 is almost 100 m above the natural timberline in the Lys valley, even though young individuals of
150 larch (*Larix decidua*) are found up to 2400 m a.s.l., evidencing an ongoing increase in timberline
151 associated with recent climate change and/or reduction in cattle grazing pressure. The whole Lys
152 proglacial area is roughly exposed to the south and only the most ancient subalpine LIA soil profiles
153 are located on the northward slopes of the 1821 morainic arc.

154 The parent glacial till is made of granitic gneiss and paragneiss belonging to the Monte Rosa nappe,
155 with minor (ca. 10%) mafic and ultramafic inclusions derived from ophiolitic outcrops in the
156 southernmost portions of the glaciated part of the massif, belonging to the Piedmontese Ophiolitic
157 Units (Mattirolo et al., 1951). Based on the observation of the stone fraction, the lithological
158 composition of the glacial till was similar in every stage of the soil chronosequences. Only the pre-
159 LIA alpine site (A6) had a higher mafic-ultramafic content (ca. 30% in weight of rock fragments)
160 than the other sites.

161 The soil moisture regime is udic (Soil Survey Staff, 2010), with a mean yearly rainfall around 1200
162 mm (Figure 2) and no dry season (alpine subatlantic climate). The south-north direction of the Lys
163 Valley increases the advection of warm, moist Mediterranean air masses from the south, increasing
164 summer rainfall, while the proximity to the main Alpine divide allows some spillover of
165 precipitation also from the north during strong foehn wind events. The mean annual air temperature
166 ranges between ca. 2°C at 1900 m a.s.l. and -1°C at 2400 m a.s.l., while winter average temperature
167 is below -4°C (Mercalli, 2003).

168

169 **2.2 Soil description vegetation survey and numerical analysis**

170 Floristic relevées were performed on 4X4 m square surfaces around the soil pits. The plant species
171 were determined according to Pignatti (1992) and single species cover (%) was visually estimated.
172 Total vegetation cover (%), proxy for plant productivity, was estimated as well.

173 Field description of soil profiles was done according to FAO (2006). Approximately 0.5-1 kg of soil
174 material was collected from every horizon in the soil pits. In the field we were not able to obtain
175 samples for the calculation of bulk densities because of excessive stoniness and/or the abundant
176 presence of medium and/or large roots. The soil chemical and physical analyses were performed
177 according to standard methods (Ministero delle Politiche Agricole e Forestali, 2000).

178 All samples were air-dried and sieved to separate the fine earth (below 2 mm) from the coarse
179 fraction. pH was measured in water (soil:water=1:2.5); total carbon (corresponding to total organic
180 carbon, TOC, thanks to the absence of carbonates) and nitrogen (TN) were analyzed by dry

181 combustion with a CN elemental analyzer (CE Instruments NA2100, Rodano, Italy). The cation
182 exchange capacity (CEC) was measured with the ammonium acetate extraction (pH 7) method, in
183 order to classify soils according to the IUSS Working Group (2006). Exchangeable base content and
184 saturation, on the ammonium acetate extracts, were measured by AAS (flame atomic absorption
185 spectrometer, Analyst 400, Perkin Elmer, Waltham, MS, USA). The particle size distribution was
186 determined by the pipette method. In order to detect the spodic properties, the oxalate and
187 dithionite-extractable fractions of Fe and Al (Fe_o, Al_o) were measured.

188 Chronofunctions of TOC concentration changes in surface horizons, of the Spodic Index (later
189 on, IS, equation 1, Soil Survey Staff, 2010) in CB-BC-Bs horizons and the ratio between the IS in
190 subsurface CB-BC-Bs horizons and in surface A-AE-E ones (later on, IS_{ratio}, equation 2) in
191 subalpine soils were calculated with the *lm* function, included in the R software (R Development
192 Core Team, 2000).

$$193 \text{ IS} = 0.5 * \text{Fe}_o + \text{Al}_o \quad (1)$$

$$194 \text{ IS}_{\text{ratio}} = \text{IS}_{(\text{B-BC})} / \text{IS}_{(\text{E-AE})} \quad (2)$$

195 As reported above, we could not calculate the soil carbon stock because no bulk density data have
196 been measured. However, we believe that the changes in concentration of the different soil
197 compounds can effectively show pedogenic trends in the studied soils, considering that the skeletal
198 fraction resulted quite constant, particularly in the first few hundred years of soil development. The
199 best variable transformation (logarithmic or power) was chosen according to the R² and the
200 significance of the regression coefficients. The chronofunctions were only descriptive, as the
201 sampling site number was excessively small to obtain statistically significant data. Moreover no
202 data were available for the 260-11000 years BP time span, and the precise ages of pre-LIA S10 and
203 A6 were not available. Significant differences in many edaphic parameters between different plant
204 covers were also checked and displayed as boxplots, using the *multcomp* R package (Hothorn et al.,
205 2008).

206

207 **3. Results**

208

209 **3.1. Vegetation primary succession**

210 **3.1.1 Alpine primary succession**

211

212 Above timberline (Table 2, Figure 3), two different grassland communities were normally found
213 outside the LIA glacial till, roughly corresponding to the lower alpine belt: flat, humid areas were

214 dominated by the *Carex curvula* association (*Caricetum curvulae*, Braun-Blanquet 1948), while
215 steeper and drier sites were characterized by *Festuca varia* (*Festucetum variae*, Braun-Blanquet,
216 1948); the observed pre-LIA site was colonized by a rather xerophilous community dominated by
217 *Festuca varia* (A5, Table 2).
218 Immediately after moraine stabilization (5-7 years after glacier retreat, P1), the pioneer species of
219 the *Epilobietum fleischeri* (quite similar to the pioneer community described by Burga et al., 2010
220 in the Morteratsch forefield) began the colonization of the raw till. Some of these species, often
221 typical of base-rich soils (Pignatti 1992), were still present on 60 years old moraines (A2, A3). Mid
222 successional species (Table 1) were common in 60-190 years old sites (A2, A3, A4); these species
223 were characteristic of disturbed, rocky and eroded soils, only weakly acidified in the surface
224 horizon. Species typical of later stages of succession appeared on 60 years old moraines (e.g., *Carex*
225 *curvula*, *Festuca varia*, *Festuca halleri*), but became more common in later stages (A4, A5). The
226 vegetation growing on 260 years old materials (A5) and the pre-LIA site (A6) was almost
227 completely devoid of early and mid-successional species. The A5 site, in particular, was covered by
228 a hygrophilous facies of the *Caricetum curvulae* (Braun-Blanquet, 1948), rich in dwarf Ericaceae
229 such as *Loiseleuria procumbens* and *Vaccinium uliginosum* subsp. *gaulterioides*, probably because
230 of microclimatic conditions favoring a long-lasting snow cover (north-west aspect).

231

232 3.1.2 Subalpine primary succession

233 Below timberline (i.e., subalpine primary succession, Table 3, Figure 3), the steep surfaces
234 deglaciaded between ca. 1950 and 1987 (not sampled) were mostly colonized by *Salix* spp.,
235 accompanied by *Rhododendron ferrugineum* and young and scattered larch trees (Treter et al.,
236 2002). An extreme vegetation patchiness existed on surfaces deglaciaded between 1922 and 1950
237 (excluding stony mounds, not considered in this study). Under an open *Larix decidua* forest,
238 surfaces covered mostly by *Rhododendron ferrugineum* (S3, S4) were intermixed with more open,
239 grazed, grass-dominated ones (SG2, SG5), which still included a few species characteristic of
240 pioneer *Epilobietum fleischeri*. Mid-successional species, such as *Salix* spp., were locally common
241 but never dominating. Species typical of subalpine forest-heath communities already showed high
242 covering rate in grassland patches, such as *Avenella flexuosa* and *Calamagrostis villosa*. Other
243 common species were typical of subalpine prairie (e.g., *Festuca varia*, *Nardus stricta*,
244 *Anthoxanthum odoratum*).

245 Surfaces deglaciaded between 1860 and ca. 1922 (S6) were covered by a quasi-climax subalpine
246 *Larix decidua* open forest, with *Rhododendron ferrugineum* dominating the understory, together
247 with *Vaccinium* spp; grasses were already characteristic of a well-developed subalpine forest

248 (*Avenella flexuosa*, *Calamagrostis villosa*, *Homogyne alpina*). The easternmost part of this area has
249 always been used for intensive cattle grazing and only scattered trees and shrubs were present
250 (SG7). Here, species typical of subalpine acidophilous grazed grassland were common, such as
251 *Nardus stricta*, *Festuca varia* and *Rhinanthus alectorolophus* (Table 3).

252 The 1821 moraine is steep and north-facing; a thick, long-lasting and creeping snow cover
253 characterizes the microclimate of this site, which is colonized mainly by green alder (*Alnus viridis*),
254 which partly inhibits high covering values by *Vaccinium myrtillus* and *Rhododendron ferrugineum*,
255 which are only locally dominant (S9).

256 The older surfaces, originated from glacial till abandoned during late glacial maxima (around 10000
257 years B.P., S10) were colonized by a climax open Larch forest, with an understory dominated by
258 *Rhododendron ferrugineum* and *Vaccinium* spp. The specific composition and covering values were
259 similar to those characterizing most of the stable forest sites in the area older than 90 years.

260 The plant cover was already around 100% in 65 years old soils, while above timberline it reached
261 the same value after 90 years.

262 **3.2 Soil properties along the chronosequences**

263 Soil development trends were well correlated with the primary succession of the vegetation cover
264 and land use; three different pathways were observed under alpine prairie, subalpine grassland and
265 typical subalpine forest-heath communities (Figure 4).

266

267 **3.2.1. Alpine soil chronosequence**

268 Above timberline (Table 4, Figure 4a), organic matter accumulation and acidification in the soil
269 surface horizons were the main pedogenetic processes in soils younger than 200 years. These
270 processes led to the formation of A horizons, with a maximum thickness and organic carbon content
271 in 260 years old soils (A5). The C/N ratio in upper alpine mineral soil horizons ranged between ca.
272 11 (A2) and 17 (A5, influenced by Ericaceae), while during the first stages of soil development
273 subsurface horizons showed much lower values. The difference between surface and subsurface
274 horizons decreased with increasing soil age, suggesting a more efficient organic matter
275 incorporation in the soil profile.

276 Weathering in subsurface horizons, releasing free Fe and Al oxi-hydroxides, led to the formation of
277 weakly developed, brownish AB and BC horizons in 190 years old soils (A4). Younger soils had
278 only A, AC and C horizons while a “true” brown, structured Bw appeared in 260 years (A5). Well-
279 developed pre-LIA soils were characterized by thick and well developed brown Bw horizons with
280 strongly acidic pH values, particularly in the A horizon.

281 Amorphous and crystalline Fe and Al oxi-hydroxides were weakly redistributed with depth
282 (increasing contents in subsurface horizon associated with a depletion in surface ones) in the 260
283 years old and in the “late glacial” soils (A5 and A6, Table 5).
284 Initial soils (P1) had near-neutral pH values and high Base Status, thanks to the abundance of
285 freshly ground, highly reactive primary minerals. Acidification and desaturation proceeded quite
286 fast, particularly in surface A horizons, together with the strong increase in organic matter. Base
287 status below 50% appeared after more than 65 years.
288 According to the WRB soil classification (IUSS Working Group, 2006) the soils up to 65 years in
289 the forefield (P1, A2, A3) were classified as Haplic Regosol (Eutric, Skeletic). After 190 years (A4),
290 weakly developed but already acidified soils were classified as Haplic Regosol (Dystric, Skeletic).
291 On the oldest LIA morainic arc and on late glacial materials (A5 and A6) Haplic Cambisol (Dystric,
292 Skeletic) were found, thanks to the presence of well developed Bw horizons (Table 4).
293 RhizoMODER humus forms characterized most of the soils older than 60 years, while A5, with an
294 ericaceous cover, had a hemiMODER (Zanella et al., 2011).
295

296 **3.2.2. Subalpine soil chronosequences**

297 Considering only stable sites below timberline, two different soil chronosequences were observed,
298 associated with subalpine forest-shrub vegetation or with subalpine grazed grassland.
299 Under subalpine grazed grassland (Figure 4b), soil processes and horizon formation were similar,
300 but faster, to the ones observed above timberline. TOC accumulation in A horizons increased with
301 age and pH values decreased to ~5.5 (SG7); in the meantime, weakly structured, brownish Bw
302 horizons formed (in 130 years old prairie sites, SG7). Where the vegetation was dominated by
303 grassland species (SG5, SG7) or by *Alnus viridis* (S8), the C/N values were below 14 (Figure 5a).
304 Under grassland vegetation, the albic horizon did not form, and the Fe and Al oxi-hydroxides did
305 not redistribute with depth.
306 Under subalpine larch forest with Rhododendron understory (Figure 4c), great morphological and
307 chemical changes characterized soils in the first 60-65 years since deglaciation (S3). Up to 6 cm of
308 litter accumulated on the soil surface, below which 3-6 cm thick, dark grayish A horizons developed
309 and weathering created yellowish BC ones (Table 6). These A horizons were characterized by the
310 absence of structure and by the juxtaposition of mineral and organic particles, creating a “salt and
311 pepper” appearance. pH values dropped from 6.5 to 5.0 in the upper horizons, while the C/N values
312 were already close to 20. The C/N values were significantly higher than under other vegetation
313 types (Figure 5a) while the pH values were lower (Figure 5b). Also the thickness of all O horizons
314 was significantly higher under forest than under grassland (Figure 5c).

315 In the following 30 years (S4, on the 1921 moraine), the A horizons were substituted by thin, visibly
316 bleached E horizons, characterized by a decrease in pedogenic Fe oxi-hydroxides (Table 7). A
317 further decrease in pH values (4.7 in the E horizon) was measured in ~130 years old soils (S6) and
318 the E horizons met the morphological requirements for the diagnostic albic horizons (Soil Survey
319 Staff, 2010 and IUSS Working Group, 2006). Under these albic horizons, weak Bs horizons formed
320 (here called CBs or BCs due to the lack of structure and the light yellowish colors; the
321 macromorphological requirements for the spodic horizon were not met, according to Soil Survey
322 Staff 2010, and IUSS Working Group, 2006). Fe redistribution was evidenced by yellowish colors
323 and by Fe_o and Fe_d depth trends (Figure 6). The E and BCs horizons further developed in 190 years
324 old soils (S8, S9), when the redistribution of Al (Al_o) became measurable. Higher Fe_o
325 concentrations in subsurface horizons was verified by the ratio between Fe_o concentrations in B-
326 BC-CB and in surface E-AE-A horizons (Figure 5d), which was significantly higher under forest-
327 shrub vegetation also in young soils. The same increasing concentration in subsurface horizons was
328 observed for Fe_d (Figure 5e). Al_o was extremely low in all LIA soils, and its redistribution with
329 depth was not significantly different under the considered vegetation types. Thus, the chemical
330 requirements (TOC > 0.5%, IS > 0.5%, ISratio >= 2) for the diagnostic spodic horizons were not
331 met. The ISratio resulted significantly higher under forest-shrub vegetation than under other land
332 covers (Figure 5f).

333 Pre-LIA climax soils were Podzols with an extremely well developed morphology, both under forest
334 and under anthropogenic grassland (not shown). These Podzols were characterized by a strong
335 illuviation of organic matter and pedogenic Fe, Al and Si oxides in the spodic horizons. Deep
336 cemented horizons (ortstein) were generally developed below the Bs (S10).

337 According to the WRB soil classification (IUSS Working Group, 2006), the soils under subalpine
338 forest/shrubs on the LIA materials, and soils under grassland up to 90 years old, were classified as
339 Haplic Regosol (Dystric, Skeletic). Subalpine “climax” soils were Ortsteinic Podzols (Skeletic)
340 (S10). Dystric Cambisols were found on 130 years old surfaces under grazed prairie (SG7).

341 Under subalpine grazed grassland, rhizoMODER humus forms were identified, characterized by the
342 presence of OF and, sometimes, OH horizons, overlying root-rich, single grain A ones. Under
343 Rhododendron-larch forest, the humus form was dysMODER in 60 years old soils (S3). It quickly
344 evolved towards MOR forms (hemiMOR in S4, S6 and S9, euMOR in S8); pre-LIA soils had
345 humiMOR forms (Zanella et al. 2011).

346

347 4. Discussion

348 4.1 Vegetation primary successions along the chronosequences

349

350 Microtopography governs the establishment of pioneer species and their turnover in the first few
351 hundred years during the primary succession (Burga et al., 2010). Considering only stable, fine
352 earth-enriched sites (as in this study), the trends in the vegetation succession are more linear.
353 As in other proglacial areas worldwide, the early stages of soil development and vegetation
354 succession were similar above and below timberline, as almost the same plant species colonized the
355 “young” morainic till, mostly belonging to the “*Epilobietum fleischeri*” (Braun Blanquet, 1948).
356 Above the treeline, the development of soils and the vegetation succession was slightly faster than
357 in other proglacial areas described in the Italian Alps (Andreis et al. 2001). The considered time
358 span (260 years) was enough for the establishment of a quasi-climax vegetation (*Curvuletum* or
359 *Festucetum variae*) while 500 years is the minimum time normally required for the establishment of
360 the climax *Curvuletum* in the Alps (Andreis et al., 2001).

361 The primary plant succession above timberline was strongly influenced by topography and
362 microclimate: for example, *Festuca varia* was common on 60 and 190 years old moraines (A3 and
363 A4) and in the climax, pre-LIA site (A6), while it was absent from the cooler and more humid 260
364 years old moraine (A5). It seems thus plausible that the further development of the vegetation cover
365 in the A3 and A4 sites will lead to the development of *Festucetum variae*, similar to “climax” A6,
366 and not towards a *Curvuletum*, similar to the one growing on the A5 site.

367 The similarity between primary vegetation successions in areas above and below timberline ends
368 soon, and already after about 60 years, larch trees (*Larix decidua*) have largely invaded the areas at
369 lower elevations, accompanied by ericaceous shrubs (e.g. *Vaccinium* spp., *Calluna vulgaris* and
370 *Rhododendron ferrugineum*) (similar to the situation described in Alaska by Boggs et al., 2010).

371 The establishment of a *Rhododendron* understory, accompanied by *Salix helvetica* and *Salix*
372 *appendiculata*, was observed earlier than in other proglacial areas in the Alps (e.g., Bernasconi et
373 al., 2011). A stabilization of the species composition below timberline was observed after the
374 establishment of ericaceous shrubs, particularly *Rhododendron ferrugineum*, on ca. 90 years old
375 surfaces. *Rhododendron ferrugineum* is the main limiting factor to the survival of pioneer species,
376 because of soil shadowing and because of the thick litter layer created by this species (Pornon and
377 Doche, 1996). A quasi-climax vegetation structure and composition was generally reached after
378 100-120 years: a much faster vegetation succession is thus observed in this work, if compared to
379 other well studied chronosequences on the Alps (e.g., Burga et al., 2010, Dümig et al., 2011).

380 The early establishment of a quasi-climax vegetation below the timberline differs from what
381 reported by Burga (1999), Burga et al. (2010) and Föllmi et al. (2009) in the Morteratsch and the
382 Rhöne proglacial areas respectively, where the most common shrub was Green alder (*Alnus viridis*),
383 with associated “megaphorbiae”, while Ericaceae (*Rhododendron ferrugineum* and *Vaccinium* spp.)
384 never reached covering values higher than 5%. The dominance by *Rhododendron ferrugineum* in
385 the Lys proglacial area was observed in sites stabilized before 1945 by Treter et al. (2002), who
386 associated this dominance to light grazing. The ecosystem mosaic, related with microtopography,
387 was particularly visible in the 1921-1945 time span, where stable sites covered by a larch-
388 *Rhododendron* forest were mixed with grazed open grassland patches and stony mounds; quasi-
389 climax subalpine shrubs (*Rhododendron ferrugineum* in particular) colonized most of the older
390 surfaces, except where grazing pressure was higher. Stony mounds (not considered in this study)
391 were common in the area younger than 91 years, and were colonized by pioneer species (Treter et
392 al., 2002).

393 Grazed sites, devoid of ericaceous shrubs and Larch trees, were colonized by different herbaceous
394 species compared to alpine grasslands, and the oldest sites had a species assemblage close to the
395 typical subalpine acidophilous *Nardus stricta* prairie. Grazing locally inhibits shrub and tree growth
396 also on “old” surfaces (SG7), but a low-pressure grazing influence on the abundance of *Ericaceae*
397 and scarcity of *Alnus viridis* and associated species can be hypothesized.

398

399 **4.2 Soil properties along the chronosequences**

400 Together with the primary vegetation succession, the speed of diagnostic horizon development and
401 taxonomic reference group change (IUSS Working Group, 2006) was faster below timberline than
402 at higher elevation. Under subalpine grazed grassland, in fact, a Haplic Cambisol (Dystric) formed
403 in 130 years (SG7), while above timberline the same taxonomic level was reached in 260 years
404 (A5). 60 years old soils below subalpine grassland were morphologically similarly developed as
405 190 years old ones above timberline (A4).

406 This higher soil development rate was probably caused by the higher productivity of ecosystems
407 below timberline, thanks to less harsh climatic conditions and by hypothetically weaker
408 cryoturbation phenomena. However, cryogenic features were not significantly more developed
409 above than below timberline, as demonstrated by the ubiquitous presence of thin and weak silt caps
410 (table 1, 3) in shallow subsurface horizons particularly in young soils (process well described by
411 Forman and Miller, 1994). Above timberline, silt caps were best developed in 60 years old soils,
412 and became less visible with increasing ages, until they disappeared in 260 years old soils. Below
413 timberline, silt caps were not evident under forest/shrub vegetation, while thin silt caps were visible

414 in shallow subsurface horizons in 60 and 90 years old soils and only in deeper C horizons of 130
415 years old soils under grazed grassland. Here they were not visible in soils older than 130 years. Silt
416 caps disappearance was probably related to the bioturbation caused by roots, which was only
417 slightly faster below than above timberline.

418 A higher ecosystem productivity below than above timberline was verified by the higher TOC
419 concentration in the fine earth of upper mineral horizons in the 130 years old soil (SG7, table 4, 6);
420 however, its trend with time was disturbed by a large variability. Above timberline, the rate of TOC
421 concentration increase declined from $0.28 \text{ g*kg}^{-1}\text{y}^{-1}$ in 70 years old soils to $0.16 \text{ g*kg}^{-1}\text{y}^{-1}$ on 260
422 years old moraines; at this point, the balance between organic carbon inputs and loss via
423 decomposition in surface horizons (steady state) was reached. Subsurface accumulation continued,
424 in quantity and depth, as shown by the increase in TOC concentration and in thickness of Bw and
425 BC-CB horizons (Table 6). Overall, under grassland the TOC concentration in surface horizons
426 tended to increase for the first few hundreds years until reaching a steady state between inputs (via
427 root and litter decay) and outputs (mineralization, erosion and leaching), according to the indicative
428 chronofunction (Figure 7a, $R^2 = 0.284$, p-value < 0.05 for all coefficients):

$$429 \quad \text{TOC} = -0.287 + 0.616 * \ln(\text{age}) \quad (3)$$

430
431
432
433 Lower TOC concentrations characterized surface mineral horizons in forest-shrub soils (Figure 7b),
434 which also showed a completely different TOC variation with time (Figure 7a). The highest TOC
435 concentration was reached in 60 years old soils. After this time, TOC concentrations started to
436 decrease, according to the descriptive chronofunction ($R^2 = 0.201$, p-value < 0.1):

$$437 \quad \text{TOC} = 1.723 - 0.09 * \ln(\text{age}) \quad (4)$$

438 The slowdown of the organic matter accumulation started quite early compared to other
439 chronosequences (e.g. Dümig et al., 2011), because of the quick establishment of the podzolization
440 process under the subalpine vegetation (He and Tang, 2008) and the development of an albic
441 horizon below the litter layer. This trend was not observed above the treeline or under subalpine
442 grazed grassland, because of limited vertical migration of soluble organic molecules and of a lower
443 organic matter production caused by the less productive vegetation. The increasing trend of TOC
444 concentration in the subsurface horizons of forest soils was faster than in the grassland ones (Figure
445 7c). Forest soils had higher TOC concentrations in subsurface horizons (Figure 7d) than grassland
446 soils, probably thanks to a vertical migration of soluble organic compounds (possibly associated to
447 a higher organic matter produced by root decay).

448 The C/N ratio was significantly related to the vegetation cover, with the highest values observed in
449 the organic horizons composed of ericaceous shrubs and Larch leaves (e.g. Boettner and Kalisz,
450 1990). Below the forest-shrub vegetation, the lowest C/N value was found under an understory

451 vegetation dominated by the N-fixing *Alnus viridis* (S8). Where the alpine prairie was enriched in
452 dwarf Ericaceae, such as *Loiseleuria procumbens* and *Vaccinium uliginosum* subsp. *gaulterioides*
453 (A5), the C/N ratio was higher. This ratio is characteristic of slowly decomposing organic matter,
454 rich in fulvic and low-molecular weight acids, and is associated to low pH values. Indeed the fastest
455 and strongest pH decrease was observed under subalpine forest/shrub vegetation, where values
456 below 5 characterized the surface mineral horizon already after 60 years since deglaciation. Under
457 grazed grassland and alpine prairie the pH reached comparable values after 130 and 260 years,
458 respectively. Since the leaching of organic acids derived from organic matter degradation is the
459 main acidifying factor in proglacial soils (Bernasconi et al. 2011), the slower pH decrease under
460 alpine vegetation could be attributed to the lower biomass production and the stronger disturbances
461 characterizing alpine habitats. However, a weak acidification was visible also in the youngest soils,
462 extremely poor in organic matter, thanks to incipient mineral weathering and leaching caused by
463 rain and snowmelt.

464

465 **4.3 Evidences of podzolization processes**

466 Both Fe_o and Al_o in the initial soil (P1) were higher in the top C1 horizon than in the underlying
467 C2. This distribution could be attributed to an initial mineral weathering caused by the organic
468 matter accumulation produced by the few plants which colonize these sites, and by oxidative
469 reactions involving fresh, reactive particles. The already detectable Fe and Al oxi-hydroxides in the
470 fresh material of the C2 horizons in the 6 years old soils could be attributed to subglacial
471 weathering processes, a well-known source of already weathered material to initial soil formation
472 processes (Anderson et al., 2000 and Dümig et al., 2011).

473 This depth trend was soon reversed: an early redistribution of pedogenic Fe oxi-hydroxides to
474 subsurface horizons was observed under subalpine larch forest with *Rhododendron* understory.
475 Higher values of Fe_o and Fe_d in subsurface BC horizons compared to the overlying A ones were
476 already observed in the ca. 60 years old soils. Despite the weaker Al_o redistribution with depth in
477 young soils, the ISratio tended to increase with time, according to the (qualitative) chronofunction
478 ($R^2 = 0.801$, p-value < 0.01 for the age coefficient, Figure 8a):

$$479 \text{ISratio} = 0.676 * \text{age}^{0.185} \quad (5).$$

480 Thus, around 350 years are needed in order to meet the ISratio requirements for the diagnostic
481 spodic horizon according to Soil Survey Staff (2010) and IUSS Working Group (2006).

482 Observing the IS and Fe_d depth trends (Figure 5 and Figure 6), the cheluviation seems to have
483 actively translocated Fe and Al oxi-hydroxides into subsurface BC and CB horizons. The IS, in

484 subsurface horizons of soils under subalpine forest-shrubs (*Rhododendron* and *Larch*), increased
485 according to the chronofunction (Figure 8b):

$$486 \text{ IS} = 0.045 + 0.003 * \text{age}^{0.8} \quad (6)$$

487 therefore, the time needed was ca. 530 years.

488 Based on the few data available, all the regressions between time since deglaciation and the spodic
489 properties had similar results: the time required to obtain a “chemical” Podzol, according to the Soil
490 Survey Staff (2010) and IUSS Working Group (2006) is around 300-600 years, which is a much
491 shorter period than the 3000 years previously calculated on the Alps (Egli et al. 2001).

492 However, these findings apply only to sites located below the timberline and covered by larch forest
493 with ericaceous understory. Below subalpine grazed grassland and below alpine prairie, Fe and Al
494 oxyhydroxides were more abundant in surface A horizons compared to subsurface BC or Bw
495 horizons. An exception was the 60 years old grassland soil (SG3), influenced by nearby larch trees,
496 and A5, thanks to abundant ericaceous dwarf shrubs (*Vaccinium uliginosum* subsp. *gaulterioides*
497 and *Loiseleuria procumbens*).

498 Similarly to other chronosequences (e.g., Burt and Alexander 1996, Egli et al. 2001), the slower
499 release and mobility of Al from the parent material, compared with Fe, could be related to the
500 presence of easily weatherable Fe-rich minerals included in the small amounts of mafic rocks in the
501 parent material. The silic minerals weathered more slowly in the considered soils, as demonstrated
502 by the Al_0/Fe_0 ratio increase with time, both in surface (not shown) and in subsurface horizons
503 (Table 7). Conversely, in more acidic parent materials, the onset of the podzolization process tended
504 to remove Al faster than Fe from the young and weakly developed E horizons (Mokma et al., 2004).
505 The E horizon formation in alpine soils has sometimes been interpreted as a stagnogley feature,
506 caused by seasonal waterlogging and alternation of reductive and oxidative conditions, common in
507 high altitude soils thanks to the abundant snow cover and high water input during the snowmelt
508 (e.g. Gensac, 1990). However, the E horizon formation under subalpine vegetation in the Lys
509 proglacial area cannot be ascribed specifically to seasonal waterlogging, as stronger seasonal
510 reductive conditions should be met above timberline, thanks to deeper winter and spring snow
511 covers. Here, E and CBs/Bs horizons were never found, and Fe/Al redistribution to subsurface
512 horizons normally was not observed. Moreover the coarse texture of the soil material should
513 facilitate drainage, decreasing the waterlogging potential in the topsoil.

514 The Fe_0/Fe_d ratio, called “activity ratio” and considered as a measure of the crystallinity of
515 pedogenic Fe oxides (Burt and Alexander, 1996), did not show any significant time trends but had
516 significantly lower values under subalpine forest-shrub vegetation than under grassland (Figure 9),
517 particularly in subsurface horizons. Alternation of reductive and oxidative conditions are known to

518 favor the crystallization of pedogenic Fe-oxi-hydroxides; however, the lower values found under
519 subalpine vegetation, particularly in subsurface BC and Bs horizons, are not explainable according
520 to surface seasonal redox conditions, for the same reason explained above. Nor the higher
521 crystallinity found in soils under forest vegetation, compared to the soils developed under even
522 more humid climates on the Alps (Dümig et al., 2011), could be easily explainable. Low Feo/Fed
523 values, however, are typical of many well developed podzolic soils in the Valle d'Aosta Region
524 (e.g. D'Amico et al., 2008).

525

526 On the Alps, Podzols are known to develop from silica-rich glacial till in late stages of pedogenesis
527 (3300 and 10000 years), as reported by Egli et al. (2001). Faster rates of E horizon formation have
528 often been measured in other boreal or mountain areas (Sauer et al., 2007). Faster Podzol
529 development in Alaska (Alexander and Burt, 1996, Ugolini, 1966), Norway (Mellor, 1986) or China
530 (He and Tang, 2008) compared to the European Alps was attributed to the maritime climate (Egli et
531 al., 2006, Dümig et al., 2011), which accelerates plant growth, soil organic matter accumulation
532 and, consequently, the rate of soil development, thanks to longer growing season and much higher
533 winter temperature. However, the establishment of mixed coniferous forests caused an immediate
534 onset of the podzolization process also in continental climates (ca. 225 years, Lichter, 1998). An
535 even faster redistribution of Fe oxi-hydroxides was observed on sandy parent materials under boreal
536 Scots pine in north-western Russia (Abakumov et al., 2010), where higher Feo contents in BCs than
537 in AE horizons were measured in 10 years old soils, and E horizons appeared in 20 years; in fact,
538 sandy parent materials are known to increase the speed of the podzolization process thanks to the
539 already weathered grain surfaces (Schaetzl and Anderson, 2005) and higher water mobility that
540 promotes vertical fluxes of elements.

541 Thus, on a global scale, the "fast" podzolization process characterizing subalpine soils developed on
542 the gneissic materials of our study area is in the "normal" ranges, but much faster than in other
543 alpine chronosequences. Similarly developed soils have been observed on the LIA morainic arcs of
544 the Miage glacier, Val Veny, in the Mont Blanc Massif and in the Verra Grande forefield, Monte
545 Rosa Group (D'Amico 2011, unpublished data). These proglacial areas in Valle d'Aosta are
546 characterized by temperature and precipitation regimes similar to other proglacial areas, in
547 particular in the northern parts of the Alps, where the podzolization processes start later and seem
548 slower than in our study area. Thus, climatic differences cannot be the cause of such a slower
549 pedogenesis on the northern side of the Alps, while the early establishment of ericaceous shrubs
550 below the treeline appeared the main driver of the early appearance of the bleached E horizon in 90
551 years old soils.

552 Plant communities on Swiss glacier forefields were dominated by alder (*Alnus viridis*), normally
553 associated with N-fixing bacterial communities (Egli et al., 2001, Burga et al., 2011), and by
554 herbaceous plant species, while *Rhododendron ferrugineum* appeared in later stages of succession
555 (Dümig et al., 2011). This vegetation creates a litter which has a weak complexing capacity, while
556 the ericaceous shrubs are able to begin a quick podzolization process in previously non-podzolic
557 soils (Bernier and Gillet 2012, 1993, Boettcher and Kalisz, 1990). This is caused by the slow
558 decomposition rates of the litter of ericaceous shrubs, due to their high amount of lignin, cellulose
559 and other recalcitrant substances, such as phenolic compounds, which reduce the soil biological
560 activity (Pornon and Doche, 1995). The litter of coniferous trees and Ericaceae produces large
561 quantities of low molecular weight and fulvic acids, which cause intense mineral weathering
562 (Schaetzl and Anderson, 2005). As humus forms are considered a good indicator of forest
563 ecosystem functioning (Michalet et al., 2001), the fast development of Mor humus forms under
564 subalpine typical *Rhododendron*-larch vegetation confirms the slow mineralization of the soil
565 organic matter, typically associated with the onset of podzolization.

566 The strong vegetation effect on pedogenesis could be enhanced by mycorrhizal fungi associated
567 with different plant species. In fact, ectomycorrhizal and ericoid (associated to Ericaceae) fungi are
568 known to increase the weathering rate in surface mineral horizons, particularly under coniferous or
569 ericaceous species, where they form mat-like structure at the boundary between the organic layer
570 and the upper mineral horizon, and extending down to the E horizons (Koele et al., 2011). These
571 fungi are able to dissolve mineral grains, extracting and chelating metals and nutrients via the
572 release of phenolic compounds, low weight organic acids, oxalate, citrate and malate (Landeweert
573 et al., 2001), which have a stronger acidifying and weathering capacity than humic molecules
574 (Ochs, 1996). Ericoid mycorrhizal fungi, in particular, produce siderophores molecules, able to
575 efficiently extract and bind Fe and other metals from primary metals (Hoffland et al., 2004).
576 Substances produced and released by ectomycorrhizal fungi, thus, increase the podzolization rate
577 under coniferous trees (Lundström et al., 2000, van Breemen et al., 2000), and even more under
578 Ericaceae, thanks to siderophore substances. Hence, also this process may contribute to the faster
579 podzolization rates found in this study.

580

581 **5. Conclusions**

582

583 In this study we characterize the main pedogenetic processes occurring in recently deglaciated areas
584 under different vegetation covers. We furthermore provide evidence of fast rates of podzolization

585 under a subalpine larch-Rhododendron forest, previously undocumented for the Alps: bleached E
586 horizons are visible in 60 years old soils, and diagnostic albic horizons appear in 120 years. In this
587 time span the cheluviation of organo-metal compounds in the underlying BC horizons was not
588 sufficient to create diagnostic spodic features, but 300-500 years seemed enough for the formation
589 of a diagnostic spodic Bs horizon and, consequently, of a “real” Podzol.
590 We suggest this relatively fast rate of podzolization be due to the specific plant community
591 succession rather than to the climatic conditions in the study area, characterized by cold
592 temperatures and significant but not exceptional precipitation amounts. The appearance of a larch-
593 Rhododendron forest could significantly influence the soil organic matter characteristics, driving
594 the soil development and the onset of soil forming processes in this proglacial area.

595

596

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605

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815 **Figure captions**

816 Fig. 1: the Lys proglacial area in the Aosta Valley Region, North-western Italy. The sampling sites
817 and the main phases of glacial retreat since 1821 are also shown.

818
819 Fig. 2: 1961-1990 average daily temperature, monthly rainfall or Snow Water Equivalent (SWE,
820 white columns) and snow on the ground (dark columns) in the Gressoney d'Eyola weather station,
821 only few km far from the studied proglacial area.

822
823 Fig. 3: conceptual diagram of vegetation development in the Lys proglacial area; the plant
824 successions observed below timberline are shown on the left, above timberline (i.e. alpine
825 successions) on the right.

826
827 Fig. 4: soil chronosequences under alpine (a), subalpine grassland (b) and typical subalpine forest-
828 heath (c).

829
830 Fig. 5: comparison between C/N ratio in surface mineral horizons (a); surface pH values (b);
831 cumulative thickness of O horizons (OL+OF+OH, c); ratio between Feo (d) and Fed (e) in
832 subsurface CB, BC, Bs and Bw horizons and in surface A, AE and E ones; ISratio (Spodicity Index
833 ratio between subsurface and surface horizons, $IS_{(B-BC)}/IS_{(E-AE)}$, f) in grassland and forest-shrub soils
834 (p-value < 0.05). In a), the forest soils low-values outlier is caused by the presence of N-fixing
835 *Alnus viridis* (S8)

836
837 Fig. 6: Fed depth trends in the LIA sites analyzed in the Lys proglacial area. Different depth trends
838 are visible for the soils under different primary successions.

839
840 Fig. 7: (a), chronofunctions of TOC concentration in surface A, AE and E horizons under grassland
841 (empty circles, continuous line) and forest-shrub vegetation (dashed line, filled squares); (b),
842 surface TOC concentration (%) in subalpine forest-shrub soils and grassland; (c), TOC
843 concentration increase with time in grassland (filled squares, continuous line) and in forest soils
844 (dashed line, empty circles); (d), almost significantly different (p-value = 0.08) TOC concentration
845 in forest-shrub compared to grassland subsoil (CB, BC, BCs, Bs and Bw horizons).

846
847 Fig. 8: (a) descriptive chronofunctions of the ISratio in forest (straight line) and grassland soils
848 (dashed line); . (b) chronofunction of IS increase in subsurface forest soil horizons.

849
850 Fig. 9: activity ratio in surface and subsurface horizons of grassland and forest soils (p-value <
851 0.05).

Table 1: Main environmental properties of the study sites; phytoclimatic level (alpine/subalpine), years of deposition of the parent material, altitude, aspect, slope steepness, land cover type and total plant cover (%) are shown.

	Phytoclimatic belt	Year of moraine deposition	Altitude	Aspect	Slope	Land cover, total plant cover (%)
P1	Alpine/subalpine	2000	2385	/	5°	Pioneer veg. (5)
A2	Alpine	1945	2420	140°	15°	Grassland (25)
A3	Alpine	1945	2415	180°	5°	Grassland (80)
A4	Alpine	1820	2460	140°	20°	Grassland (100)
A5	Alpine	1755	2480	280°	30°	Grassland / dwarf shrubs (100)
A6	Alpine	ca. 11500 BP	2510	220°	30°	Grassland (100)
SG2	Subalpine	1945	2180	200°	5°	Grazed grassland (100)
S3	Subalpine	1945	2180	220°	5°	Larch-rhododendron Forest (100)
S4	Subalpine	1920	2155	340°	2°	Larch-rhododendron Forest (100)
SG5	Subalpine	1920	2155	160°	2°	Grazed grassland (80)
S6	Subalpine	1880	2105	180°	2°	Larch-rhododendron Forest (100)
SG7	Subalpine	1880	2138	160°	2°	Grazed grassland (100)
S8	Subalpine	1821	2005	20°	30°	Larch-rhododendron Forest (100)
S9	Subalpine	1821	2005	0°	30°	Larch-rhododendron Forest (100)
S10	Subalpine	ca. 11500 BP	1995	90°	20°	Larch-rhododendron Forest (100)

Table 2: Alpine primary succession; increasing age of the site is from left to right; the % cover of each species (based on visual estimation) is shown in each site. The differentiation between early successional, mid successional, late successional and ubiquitous species is based on Pignatti (1992) and on the Landolt indices (Landolt, 1977): on acidic parent materials, basophilous species should be restricted on weakly developed, initial soils, not yet acidified and desaturated.

	P1	A2	A3	A4	A5	A6
Early successional						
<i>Cerastium uniflorum</i>	1	3	0	0	0	0
<i>Epilobium fleischeri</i>	1	5	2	0	0	0
<i>Linaria alpina</i>	1	1				
<i>Oxyria digyna</i>	1	0	0	0	0	0
<i>Saxifraga oppositifolia</i>	0	0	1	0	0	0
<i>Trisetum distichophyllum</i>	1	3	0	0	0	0
Mid-successional						
<i>Achillea nana</i>	0	5	0	5	0	0
<i>Agrostis schraderiana</i>	1	0	0	0	0	0
<i>Anthyllis vulneraria</i> subsp. <i>valesiaca</i>	0	2	2	0	0	0
<i>Aster alpinus</i>	0	1	1	0	0	0
<i>Campanula cochleariifolia</i>	0	0	0	1	0	0
<i>Carex norvegica</i>	0	0	1	0	0	0
<i>Carex ornithopodioides</i>	0	0	1	0	0	0
<i>Erigeron alpinus</i>	0	1	1	0	0	0
<i>Gentiana nivalis</i>	0	0	1	0	0	0
<i>Luzula spicata</i>	0	0	2	3	0	0
<i>Myosotis alpestris</i>	0	0	10	10	0	0
<i>Saxifraga exarata</i>	1	0	2	0	0	0
<i>Saxifraga paniculata</i>	0	0	2	0	0	0
<i>Silene excapa</i>	1	0	5	1	0	0
<i>Silene rupestris</i>	0	0	0	1	0	0
<i>Trifolium pallescens</i>	0	30	10	5	1	0
Late successional						
<i>Antennaria dioica</i>	0	0	0	0	1	0
<i>Anthoxanthum alpinum</i>	0	0	1	0	0	0
<i>Bellis perennis</i>	0	0	0	0	1	0
<i>Botrychium lunaria</i>	0	0	0	1	0	0
<i>Carex curvula</i>	0	0	20	5	5	0
<i>Carex sempervirens</i>	0	0	0	0	0	5
<i>Coeloglossum viride</i>	0	0	0	0	1	0
<i>Euphrasia rohoskoviana</i>	0	0	3	0	0	0
<i>Festuca halleri</i>	0	1	3	5	2	2

<i>Festuca varia</i>	0	1	10	20	0	50
<i>Galium anysophyllon</i>	0	0	0	5	0	0
<i>Gentiana acaulis</i> subsp. <i>Koch</i>	0	0	0	1	0	1
<i>Geum montanum</i>	0	0	0	5	0	10
<i>Hieracium piloselloides</i>	0	0	0	0	1	0
<i>Homogyne alpina</i>	0	0	0	0	1	0
<i>Hupertia selago</i>	0	0	0	1	1	0
<i>Juncus trifidus</i>	0	0	10	0	0	0
<i>Juniperus communis</i>	0	0	5	0	0	0
<i>Loiseleuria procumbens</i>	0	0	0	0	40	0
<i>Lotus corniculatus</i> subsp. <i>alpinus</i>	0	0	5	3	0	0
<i>Luzula alpinopilosa</i>	0	0	0	0	2	1
<i>Luzula lutea</i>	0	0	0	0	1	0
<i>Minuartia recurva</i>	0	5	1	0	0	0
<i>Nardus stricta</i>	0	0	0	0	0	5
<i>Pedicularis kernerii</i>	0	0	0	1	1	0
<i>Pedicularis verticillata</i>	0	0	0	1	0	0
<i>Poa nemoralis</i>	0	0	2	0	0	0
<i>Pulsatilla alpina</i>	0	0	0	1	0	1
<i>Rhinanthus alectorolophus</i>	0	0	1	0	0	0
<i>Soldanella verna</i>	0	0	0	0	0	10
<i>Trichophorum caespitosum</i>	1	0	0	0	3	15
<i>Trifolium alpinum</i>	0	0	5	1	5	10
<i>Vaccinium uliginosum</i> subsp. <i>gaulterioider</i>	0	0	20	0	60	0
<i>Valeriana celtica</i>	0	0	0	0	2	0
<i>Veronica aphylla</i>	0	0	1	0	0	2
Ubiquitous						
<i>Agrostis rupestris</i>	0	0	5	0	1	2
<i>Bartsia alpina</i>	0	0	1	0	2	0
<i>Cerastium cerastioides</i>	0	0	1	0	0	10
<i>Cirsium spinosissimum</i>	0	0	1	1	0	1
<i>Festuca rubra</i>	0	1	0	0	0	0
<i>Festuca violacea</i>	0	0	5	0	0	0
<i>Leontodon helveticus</i>	0	0	1	2	0	0
<i>Leucanthemopsis alpina</i>	0	0	0	5	0	0
<i>Minuartia sedoides</i>	0	0	2	0	0	0
<i>Poa alpina</i>	0	2	5	5	2	0
<i>Polygonum viviparum</i>	0	0	2	0	5	0
<i>Potentilla frigida</i>	0	0	0	3	0	0
<i>Primula hirsuta</i>	0	0	0	0	1	0
<i>Rhododendron ferrugineum</i>	0	0	0	0	5	0
<i>Salix helvetica</i>	0	0	5	0	5	0
<i>Saxifraga moschata</i>	0	0	1	0	0	0
<i>Sempervivum arachnoideum</i>	0	10	2	1	0	0

<i>Sempervivum montanum</i>	0	3	3	2	0	0
<i>Veronica bellidioides</i>	0	0	1	0	0	0
<i>Veronica fruticans</i>	0	10	0	0	0	0

<i>Cirsium spinosissimum</i>	0	0	0	0	0	0	0	0	1	0
<i>Erigeron alpinus</i>	0	0	0	0	0	0	0	0	1	0
<i>Festuca varia</i>	0	20	15	0	10	0	0	0	30	0
<i>Festuca violacea</i>	0	5	0	0	0	0	0	0	15	0
<i>Galium album</i>	0	10	0	0	0	0	0	0	0	0
<i>Galium anisophyllum</i>	0	5	0	0	0	0	0	0	0	0
<i>Geum montanum</i>	0	1	0	0	0	0	0	1	5	0
<i>Helianthemum nummularium</i> subsp. <i>grandiflorum</i>	0	0	0	0	0	0	0	0	1	0
<i>Hieracium pilosella</i>	0	0	0	0	0	0	0	0	5	0
<i>Hieracium prenanthes</i>	0	0	0	0	0	0	0	1	0	0
<i>Knautia arvensis</i>	0	0	0	0	0	0	0	0	1	0
<i>Leucanthemum vulgare</i>	0	0	0	0	0	0	0	0	1	0
<i>Lotus corniculatus</i> aggr.	0	0	5	0	1	0	0	1	5	0
<i>Nardus stricta</i>	0	0	0	0	5	20	0	0	20	0
<i>Nigritella nigra</i>	0	0	0	0	0	0	0	0	1	0
<i>Phyteuma orbicularis</i>	0	1	0	0	0	0	0	0	0	0
<i>Plantago alpina</i>	0	0	0	0	0	0	0	0	1	0
<i>Poa supina</i>	0	2	0	0	0	0	0	0	0	0
<i>Polygonum viviparum</i>	0	1	0	0	0	0	0	1	0	0
<i>Rhinanthus alectorolophus</i>	0	0	5	0	0	0	0	0	5	0
<i>Schoenus nigricans</i>	0	2	0	0	0	0	0	0	0	0
<i>Sempervivum montanum</i>	0	0	2	0	0	0	0	0	0	0
<i>Solidago virgaurea</i> subsp. <i>minor</i>	0	0	0	0	0	0	0	1	0	0
<i>Thymus serpyllum</i>	0	0	0	0	0	0	0	0	5	0
<i>Trichophorum caespitosum</i>	1	5	0	0	0	0	0	0	3	0
<i>Trifolium pratense</i> aggr.	0	1	10	0	0	0	0	0	3	0
<i>Trollius europaeus</i>	0	0	0	0	0	0	0	1	0	0
Subalpine forest– climax (selection)										
<i>Alnus viridis</i>	0	30	0	0	0	0	30	1	0	10

Tab. 4: Macromorphological properties of the soils in the alpine chronosequence. Structure: GR = granular; PL = platy; PS = subangular polyhedral; MA = massive; RS = rock structure; AB = absent; M = matted (O horizons). The number indicates the size class of the soil structure: 1 = very fine, 2 = fine, 3 = medium, 4 = coarse; Consistence: LO loose; FR friable; FI firm; VFI very firm; The second number symbolizes the strength of the aggregates: 1 = very weak, 2 = weak, 3 = moderate, 4 = strong, 5 = very strong. Silt caps and their thickness are shown as well (1= up to 1 mm thick, visible on few rock fragments, and 2 = up to 2 mm thick and visible on many rock fragments).

Profile	Year of deposition	Horizon	Depth (cm)	Munsell Color (moist)	Sand (%)	Clay (%)	Structure (or cementation)	Consistence (moist)	Rock fragments (volume, %)	Siltcaps
P1	2004	C1	0-8	5Y5.5/3	69.1	2.1	PL3/AB	LO	60	
		C2	8-47+	5Y6/1	80.3	0.8	PL3/AB	LO	60	1
A2	c. 1950	AC	0-7	2.5Y4/3	82.3	1.1	AB	LO	50	
		C1	7-16	2.5Y4/4	83.4	1.1	AB	LO	60	1
A3	c. 1950	C2	16-35+	2.5Y5/3	81.5	1.0	AB	LO	60	2
		OL	0-0.5							
A4	1821	A	0-8	10YR4/3	81.2	1.8	GR2	LO1	50	
		AC	8-25	5Y4/2	81.8	1.5	AB	LO	70	
		C	25-48+	5Y4/1	82.9	1.5	AB	LO	70	2
A5	c. 1755	OF	0-3							
		A	3-14	10YR3/2	78.5	2.1	M	LO2	40	
		AB	14-17	10YR4/4.5	79.2	2.0	SP1	LO2	60	
		BC	17-37	10YR4/3	81.2	1.6	AB	LO	50	
		C	37-45+	2.5Y5/4	80.6	1.4	AB	LO	50	1
A6	c. 11000 BP	OL	0-1							
		OF	1-3	7.5YR2/1						
		A1	3-10	7.5YR2.5/1	79.5	2.6	M, PL3	FR3	20	
		A2	10-22	7.5YR3/2	81.3	1.9	SP2	FR2	20	
		Bw	22-40	10YR5/4	84.6	1.1	SP2	LO1	30	
A6	c. 11000 BP	C	40-50+	2.5Y5/4	94.2	0.5	AB	LO	30	
		OL-OF	0-6							
		A	6-12/16	7.5YR2.5/2	69.2	5.2	GR/GM2	LO2	5	
		BA	12/16-48	7.5YR3/3	73.0	4.7	GR2	LO2	40	
		Bw	48-72	8YR3/6	84.1	3.5	GR2	FR2	30	
IIBC	72-106+	10YR3/6	82.1	1.7	GR1	FR1	10			

Tab. 5: Chemical properties of the alpine soil chronosequences.

	Horizon	pH	C	Feo	Alo	Fed	Feo/Fed
			%	g/kg	g/kg	g/kg	
P1	C1	6.6	0.04	0.87	0.21	1.87	0.47
	C2	6.8	0.00	0.80	0.18	1.50	0.53
A2	AC	5.9	1.71	1.00	0.22	1.80	0.56
	C1	5.2	0.51	1.20	0.19	1.90	0.63
A3	C2	5.6	0.17	0.80	0.18	1.40	0.57
	A	6.1	1.46	1.35	0.44	3.15	0.44
	AC	6.4	0.33	1.23	0.26	2.60	0.46
A4	C	6.5	0.07	1.73	0.21	2.15	0.77
	O	5.1	28.00				
	A	4.7	1.90	1.80	0.70	3.40	0.53
	AB	5	0.93	1.60	0.70	3.50	0.46
	BC	5	0.50	1.30	0.50	2.80	0.46
A5	C	5.1	0.31	1.30	0.40	2.80	0.46
	A1	4.5	4.14	1.80	0.90	4.00	0.45
	A2	4.9	2.53	3.00	1.60	8.20	0.37
	Bw	5.4	0.55	1.70	1.10	8.90	0.19
A6	OL	4.6	23.24				
	A	4.6	4.83	7.25	2.46	16.20	0.45
	BA	5.4	2.19	7.70	5.68	14.80	0.52
	Bw	5.6	1.45	4.65	4.11	17.10	0.27
	BC	5.7	0.77	4.28	3.58	16.40	0.26

Tab. 6: Macromorphological properties of the soils in the subalpine chronosequence. Structure: GR = granular; PL = platy; PS = subangular polyhedral; MA = massive; RS = rock structure; AB = absent; M = matted (O horizons). The number indicates the size class of the soil structure: 1 = very fine, 2 = fine, 3 = medium, 4 = coarse; Consistence: LO loose; FR friable; FI firm; VFI very firm; The second number symbolizes the strength of the aggregates: 1 = very weak, 2 = weak, 3 = moderate, 4 = strong, 5 = very strong. Silt caps and their thickness are shown as well (1= up to 1 mm thick, visible on few rock fragments, and 2 = up to 2 mm thick and visible on many rock fragments, 3 = up to 2 mm thick and visible on most rock fragments).

Profile	Year of deposition	Hor.	Depth (cm)	Munsell Color (moist)	Sand (%)	Clay (%)	Structure (or cementation)	Consistence (moist)	Rock fragments (volume, %)	Silt caps
P1	2004	C1	0-8	5Y5.5/3	69.1	2.1	PL3/AB	LO	60	
		C2	8-47+	5Y6/1	80.3	0.8	PL3/AB	LO	60	1
SG2	c. 1950	OL	0-1							
		OH	1-3	7.5YR2/2						
		A	3-6.5	10YR4/2	81.3	1.1	GR1	LO	40	
		CB	6.5-18	10YR5/4	82.3	1.1	GR1	LO	50	
S3	c. 1950	C	18-40+	2.5Y5/3	86.4	0.8	M	M	50	3
		OL	0-1							
S4	1921	OF	1-3	7.5YR2/2						
		OH	3-6	7.5YR2/2			M	3		
		A	6-12	2.5Y4/1.5	78.6	1.9	GR1	LO	30	
		CB	12-25+	1Y5/4	81.3	1.1	AB	LO	80	
		OL	0-1							
SG5	1921	OF	1-2	5YR2/2						
		OH	2-6	5YR2/2						
		E	6-9	5YR5/1	78.3	2.1	GR1	LO	40	
		CBs/C	9-37+	10YR5/4, 10YR6/3	81.7	1.6	AB	LO	60	
		A	4-17	10YR4/3	80.5	1.7	GR1	LO	60	
S6	c. 1880	C	17-30+	10YR5/3	81.7	0.8	AB	LO	80	1
		OL	0-1							
		OF-OH	1-5	5YR2.5/2			M	3		
		E	5-7.5	10YR6/2	68.7	2.2	GR1	M	50	
SG7	c. 1880	BCs	7.5-21	10YR4.5/4	74.5	1.1	PS2	FR1	70	
		C	21-27+	10YR5/3	75.6	0.9	AB	LO	70	
		OH	0-1							
S8	1821	A	1-3	7.5YR2/1	76.5	3.1	M	3	20	
		Bw	3-20/30	10YR4/5	78.3	2.2	PS2	FR1	70	
		CB	20/30-30/38	10YR5/3	80.2	1.1	PS1	LO	50	1
		C	30/38-45	2.5Y5/3	81.6	1.0	AB	Lo	60	3

		OF	2-4	7.5YR2.5/2				M1	LO	
		E	4-7	10YR5.5/2, 10YR3/1	67.5	2.1		GR1	LO1	50
		BCs	7-21	10YR5/4	81.9	1.0		GR1	LO1	60
		C	21-42	10YR5/3	82.6	0.8		AB (PS1)	LO1	80
S9	1821	OL	0-1							
		OF	1-5	5YR2/2						
		OH	5-10/7	5YR2/2						
		AE	10/7-12	10YR4/3	68.2	1.9		GR0	M	30
		E	12-18	10YR5/2	66.5	2.1		GR1	M	40
		BCs	18-30	10YR4/4	79.0	2.0		PS2	FR1	40
		C	30-40+	10YR5/3	84.2	0.9		AB	LO	50
S10	c. 11000 BP	OL-OF	0-3							
		OH	3-21/12	7.5YR2/1				GR1	LO	
		AE/OH	21/12- 20/30	7.5YR2/2	59.2	8.6		AB	MA3	0
		Eh	20/30- 34/46	10YR5/3	64.4	7.6		AB/PL2	LO	60
		EBh	34/46-54	7.5YR4/4	66.3	7.5		PL2	LO	60
		Bhs	54-68	5YR4/4	77.8	6.1		PS2	LO2	60
		Bsm1	68-87	5YR4/4	79.2	5.9		Cemented	M	80
		Bsm2	87-120	6YR5/8	81.3	5.1		Cemented	M	80
		CBm	120-135	10YR5/8	83.4	4.9		Strongly cemented	M	80
		C(m)	135-150+	2.5Y5/6	79.6	4.8		Partly cemented	M2/LO	80

Tab. 7: Chemical properties of the subalpine soils along the Lys forefield chronosequences.

	Horizon	pH	C	Feo	Alo	Fed	Feo/Fed	Alo/Feo
			%	%	%	%		
P1	C1	6.4	0.04	0.09	0.20	0.19	0.47	0.22
	C2	6.6	0	0.08	0.18	0.15	0.53	0.25
SG2	OH	5.9	26.58					
	A	5.8	1.5	0.11	0.27	0.19	0.58	0.27
	CB	5.8	0.44	0.14	0.36	0.25	0.56	0.29
S3	C	5.9	0.15	0.10	0.31	0.21	0.48	0.30
	OL/OF	5.6	26.89					
	A	5	1.92	0.07	0.19	0.18	0.39	0.29
S4	CB	5.2	0.68	0.11	0.31	0.23	0.48	0.27
	OL/OF	5.1	41.2					
	OH	4.9	18.44					
SG5	E	5.1	1.24	0.07	0.18	0.32	0.22	0.29
	CBs	5.2	0.34	0.14	0.22	0.42	0.33	0.14
	OF	5.8	14.32					
S6	A	5.5	0.55	0.17	0.44	0.38	0.45	0.24
	C	5.8	0.19	0.12	0.51	0.16	0.75	0.50
	OF-OH	4.4	35.24					
SG7	E	4.7	1.12	0.07	0.49	0.16	0.44	0.71
	BCs	5.1	0.859	0.17	0.58	0.40	0.43	0.35
	C	5.3	0.51	0.15	0.71	0.35	0.43	0.47
S8	OH	5.6	11.78					
	A	5.6	8.18	0.15	0.61	0.28	0.54	0.40
	Bw	5	0.74	0.15	0.43	0.27	0.56	0.27
S9	CB	5.7	0.33	0.13	0.34	0.24	0.54	0.23
	C	5.4	0.14	0.09	0.22	0.18	0.5	0.22
	OL-OF	4.6	16.32					
S10	E	5	1.19	0.11	0.47	0.17	0.65	0.45
	BCs	5.1	0.75	0.17	0.80	0.67	0.26	0.47
	C	5.2	0.31	0.15	0.48	0.29	0.52	0.33
S9	OH/OF	5	18.72					
	E	4.5	0.8	0.11	0.38	0.24	0.46	0.36
	Bs	4.7	0.76	0.20	0.74	0.41	0.49	0.35
S10	OL-OF	3.6	45.21					
	OH	3.5	38.53					
	AE/OH	3.6	5.4	0.12	0.15	0.38	0.32	1.17
	E	4.1	1.02	0.09	0.09	0.23	0.41	1.00
	EBh	4.2	1.1	0.45	0.13	0.77	0.58	0.29
	Bs	4.9	2.63	1.13	0.77	4.06	0.28	0.68
	Bsm1	5.3	2.31	0.76	0.82	4.02	0.19	1.08
	Bsm2	5.4	0.86	0.29	0.73	1.56	0.19	2.52
	CBm	5.4	0.3	0.11	0.53	0.81	0.14	4.82
C(m)	6	0.11	0.11	0.34	0.44	0.25	3.00	

















