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

DOI:10.1016/j.catena.2013.10.001

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

Michele E. D'Amico*, Michele Freppaz, Gianluca Filippa, , Ermanno Zanini

DISAFA, Università degli Studi di Torino, Via Leonardo da Vinci 44, Grugliasco (TO), Italy and NatRisk

*: corresponding author. tel: +39 3490611313; e-mail: ecomike77@gmail.com

Abstract

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

In order to increase our knowledge on the effects of time and vegetation primary succession on soil development in proglacial areas, we sampled soils and surveyed plant communities on stable points in the proglacial area of the Lys glacier, in the Italian north-western Alps (Valle d'Aosta Region). The sampling points were located on dated sites (based on literature and/or historical photographs). Glacial till is attacked by weathering processes immediately after deposition and stabilization, with a consequent loss of soluble compounds, decrease of pH and primary mineral weathering. The speed of these processes was largely increased after the establishment of a continuous vegetation cover, thanks to the organic matter accumulation caused by litter input and root decomposition below the soil surface.

On sialic glacial tills, below timberline and under a quasi-climax Larch - Rhododendron forest, a fast and steady decrease in pH values, increase in organic matter content and horizon differentiation was observed. In particular, genetic eluvial horizons formed in approximately 60 years, while diagnostic albic horizons were developed after ca. 90 years, pointing to an early start of the podzolization processes. Cheluviation of Fe and, secondarily, Al were analytically verified. However, illuviation of Fe, Al and organic matter in incipient B horizons were not sufficient to obtain diagnostic spodic horizons on LIA materials.

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

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

Keywords

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

Highlights

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

- Vegetation seemed very important in the early podzolization onset.

1. **Introduction**

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

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

In the fifties some works described Podzol-like soils on recently deglaciated surfaces also in the Alps: for example, Jenny (1958) showed a 315 years old soil which had a 11 cm thick light gray horizon overlying a 14 cm thick brownish one, under subalpine ericaceous shrubs in the Rhone Gletscher forefield. However, in the alpine range, much longer periods since deglaciation were normally required for the establishment of the podzolization process than in other boreal habitats. Under subalpine vegetation, the youngest soils were usually Lithic and Skeletic Leptosols; after 120 years soil evolution led to the development of Humi-Skeletic Leptosols or, locally, to weakly developed Dystric Cambisols (Righi et al., 1999, Egli et al., 2001, Egli et al., 2006). A continuous humus layer appeared in the oldest soils, together with weak signs of chemical weathering (formation of Fe and Al oxyhydroxides). Oxalate extractable Fe and Al decreased with depth, and no cheluviation was visible (Egli et al., 2006). In the Alps, Dystric Cambisols are normally found on 250-300 years old surfaces, while Podzols appear after 600 years in very humid sites (1800 mm/y, Zech and Wilcke, 1977), but usually more than 1300 years are needed for the development of E horizons (Egli et al., 2003). In central Switzerland, in areas characterized by mean annual precipitation equal to ~2000 mm/y, Podzols are found in 3300 years old sites (Egli et al., 2001, Dümig et al., 2011), but they are normally weakly developed (Righi et al., 1999).

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

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

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

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

2 Material and Methods

2.1 Study area

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

We sampled a recent soil (P1, ca. 7-10 years old) as representative of fresh, raw till (starting point of soil development and vegetation succession for both alpine and subalpine chronosequences). Above timberline (“alpine” chronosequence), we sampled soils formed at 4 different sites under common vegetation covers (Table 1). Below timberline (subalpine chronosequence), we sampled soils at 5 different sites (Table 1); if two different vegetation types occurred on same-age surfaces, we opened a soil pit under each type (i.e. S3, S4, S6 were observed under larch forest with Rhododendron ferrugineum, SG2, SG5, SG7 below grazed grassland). S10 and A6 are on late-glacial (Younger Dryas) till (more than 10000-11000 years old, according to Pelfini et al., 1997). This age is attributed thanks to similarities with most other proglacial areas in the Alps (Ivy-Ochs et al., 2009). We excluded areas visibly disturbed by erosion or deposition processes. All soil profiles were chosen as representative among a much larger number of soil observations by soil coring. Since small scale topographic differences have strong effects on soil development and vegetation succession dynamics (Burga et al. 2010), sites with similar surface rockiness and slope were chosen.

Unfortunately, below the present-day glacier front the slope is steep and eroded (“Rocce di Salzen”), and it becomes reasonably stable only 200 m below, where the material was left by the glacier around year 1945. Therefore, no soil has been sampled and analyzed with ages ranging between 5/10 and 60/70 years.

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

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

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

2.2 Soil description vegetation survey and numerical analysis

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

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

All samples were air-dried and sieved to separate the fine earth (below 2 mm) from the coarse fraction. pH was measured in water (soil:water=1:2.5); total carbon (corresponding to total organic carbon, TOC, thanks to the absence of carbonates) and nitrogen (TN) were analyzed by dry combustion with a CN elemental analyzer (CE Instruments NA2100, Rodano, Italy). The cation exchange capacity (CEC) was measured with the ammonium acetate extraction (pH 7) method, in order to classify soils according to the IUSS Working Group (2006). Exchangeable base content and saturation, on the ammonium acetate extracts, were measured by AAS (flame atomic absorption spectrometer, Analyst 400, Perkin Elmer, Waltham, MS, USA). The particle size distribution was determined by the pipette method. In order to detect the spodic properties, the oxalate and dithionite-extractable fractions of Fe and Al (Feo, Alo) were measured.

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

$$IS = 0.5*Feo + Alo \qquad (1)$$

$$ISratio = IS_{(B\text{-}BC)}/IS_{(E\text{-}AE)} \qquad (2)$$

As reported above, we could not calculate the soil carbon stock because no bulk density data have been measured. However, we believe that the changes in concentration of the different soil compounds can effectively show pedogenic trends in the studied soils, considering that the skeletal fraction resulted quite constant, particularly in the first few hundred years of soil development. The best variable transformation (logarithmic or power) was chosen according to the R^2 and the significance of the regression coefficients. The chronofunctions were only descriptive, as the sampling site number was excessively small to obtain statistically significant data. Moreover no data were available for the 260-11000 years BP time span, and the precise ages of pre-LIA S10 and A6 were not available. Significant differences in many edaphic parameters between different plant covers were also checked and displayed as boxplots, using the *multcomp* R package (Hothorn et al., 2008).

3. Results

3.1. Vegetation primary succession

3.1.1 Alpine primary succession

Above timberline (Table 2, Figure 3), two different grassland communities were normally found outside the LIA glacial till, roughly corresponding to the lower alpine belt: flat, humid areas were dominated by the *Carex curvula* association (*Caricetum curvulae*, Braun-Blanquet 1948), while steeper and drier sites were characterized by *Festuca varia* (*Festucetum variae*, Braun-Blanquet, 1948); the observed pre-LIA site was colonized by a rather xerophilous community dominated by *Festuca varia* (A5, Table 2).

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

3.1.2 Subalpine primary succession

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

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

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

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

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

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

3.2 Soil properties along the chronosequences

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

3.2.1. Alpine soil chronosequence

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

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

Amorphous and crystalline Fe and Al oxi-hydroxides were weakly redistributed with depth (increasing contents in subsurface horizon associated with a depletion in surface ones) in the 260 years old and in the “late glacial” soils (A5 and A6, Table 5).

Initial soils (P1) had near-neutral pH values and high Base Status, thanks to the abundance of freshly ground, highly reactive primary minerals. Acidification and desaturation proceeded quite fast, particularly in surface A horizons, together with the strong increase in organic matter. Base status below 50% appeared after more than 65 years.

According to the WRB soil classification (IUSS Working Group, 2006) the soils up to 65 years in the forefield (P1, A2, A3) were classified as Haplic Regosol (Eutric, Skeletic). After 190 years (A4), weakly developed but already acidified soils were classified as Haplic Regosol (Dystric, Skeletic). On the oldest LIA morainic arc and on late glacial materials (A5 and A6) Haplic Cambisol (Dystric, Skeletic) were found, thanks to the presence of well developed Bw horizons (Table 4).

RhizoMODER humus forms characterized most of the soils older than 60 years, while A5, with an ericaceous cover, had a hemiMODER (Zanella et al., 2011).

3.2.2. Subalpine soil chronosequences

Considering only stable sites below timberline, two different soil chronosequences were observed, associated with subalpine forest-shrub vegetation or with subalpine grazed grassland.

Under subalpine grazed grassland (Figure 4b), soil processes and horizon formation were similar, but faster, to the ones observed above timberline. TOC accumulation in A horizons increased with age and pH values decreased to ~5.5 (SG7); in the meantime, weakly structured, brownish Bw horizons formed (in 130 years old prairie sites, SG7). Where the vegetation was dominated by grassland species (SG5, SG7) or by *Alnus viridis* (S8), the C/N values were below 14 (Figure 5a). Under grassland vegetation, the albic horizon did not form, and the Fe and Al oxi-hydroxides did not redistribute with depth.

Under subalpine larch forest with Rhododendron understory (Figure 4c), great morphological and chemical changes characterized soils in the first 60-65 years since deglaciation (S3). Up to 6 cm of litter accumulated on the soil surface, below which 3-6 cm thick, dark grayish A horizons developed and weathering created yellowish BC ones (Table 6). These A horizons were characterized by the absence of structure and by the juxtaposition of mineral and organic particles, creating a “salt and pepper” appearance. pH values dropped from 6.5 to 5.0 in the upper horizons, while the C/N values were already close to 20. The C/N values were significantly higher than under other vegetation types (Figure 5a) while the pH values were lower (Figure 5b). Also the thickness of all O horizons was significantly higher under forest than under grassland (Figure 5c).

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

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

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

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

4. Discussion

4.1 Vegetation primary successions along the chronosequences

Microtopography governs the establishment of pioneer species and their turnover in the first few hundred years during the primary succession (Burga et al., 2010). Considering only stable, fine earth-enriched sites (as in this study), the trends in the vegetation succession are more linear.
As in other proglacial areas worldwide, the early stages of soil development and vegetation succession were similar above and below timberline, as almost the same plant species colonized the "young" morainic till, mostly belonging to the "*Epilobietum fleischeri*" (Braun Blanquet, 1948).
Above the treeline, the development of soils and the vegetation succession was slightly faster than in other proglacial areas described in the Italian Alps (Andreis et al. 2001). The considered time span (260 years) was enough for the establishment of a quasi-climax vegetation (*Curvuletum* or *Festucetum variae*) while 500 years is the minimum time normally required for the establishment of the climax *Curvuletum* in the Alps (Andreis et al., 2001).
The primary plant succession above timberline was strongly influenced by topography and microclimate: for example, *Festuca varia* was common on 60 and 190 years old moraines (A3 and A4) and in the climax, pre-LIA site (A6), while it was absent from the cooler and more humid 260 years old moraine (A5). It seems thus plausible that the further development of the vegetation cover in the A3 and A4 sites will lead to the development of *Festucetum variae*, similar to "climax" A6, and not towards a *Curvuletum*, similar to the one growing on the A5 site.
The similarity between primary vegetation successions in areas above and below timberline ends soon, and already after about 60 years, larch trees (*Larix decidua*) have largely invaded the areas at lower elevations, accompanied by ericaceous shrubs (e.g. *Vaccinium* spp., *Calluna vulgaris* and *Rhododendron ferrugineum*) (similar to the situation described in Alaska by Boggs et al., 2010).
The establishment of a *Rhododendron* understory, accompanied by *Salix helvetica* and *Salix appendiculata*, was observed earlier than in other proglacial areas in the Alps (e.g., Bernasconi et al., 2011). A stabilization of the species composition below timberline was observed after the establishment of ericaceous shrubs, particularly *Rhododendron ferrugineum*, on ca. 90 years old surfaces. *Rhododendron ferrugineum* is the main limiting factor to the survival of pioneer species, because of soil shadowing and because of the thick litter layer created by this species (Pornon and Doche, 1996). A quasi-climax vegetation structure and composition was generally reached after 100-120 years: a much faster vegetation succession is thus observed in this work, if compared to other well studied chronosequences on the Alps (e.g., Burga et al., 2010, Dümig et al., 2011).

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

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

4.2 Soil properties along the chronosequences

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

This higher soil development rate was probably caused by the higher productivity of ecosystems below timberline, thanks to less harsh climatic conditions and by hypothetically weaker cryoturbation phenomena. However, cryogenic features were not significantly more developed above than below timberline, as demonstrated by the ubiquitous presence of thin and weak silt caps (table 1, 3) in shallow subsurface horizons particularly in young soils (process well described by Forman and Miller, 1994). Above timberline, silt caps were best developed in 60 years old soils, and became less visible with increasing ages, until they disappeared in 260 years old soils. Below timberline, silt caps were not evident under forest/shrub vegetation, while thin silt caps were visible in shallow subsurface horizons in 60 and 90 years old soils and only in deeper C horizons of 130 years old soils under grazed grassland. Here they were not visible in soils older than 130 years. Silt caps disappearance was probably related to the bioturbation caused by roots, which was only slightly faster below than above timberline.

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

$$TOC = -0.287 + 0.616*\ln(age) \qquad (3)$$

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

$$TOC = 1.723 - 0.09 * \ln(age) \qquad (4)$$

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

The C/N ratio was significantly related to the vegetation cover, with the highest values observed in the organic horizons composed of ericaceous shrubs and Larch leaves (e.g. Boettner and Kalisz, 1990). Below the forest-shrub vegetation, the lowest C/N value was found under an understory vegetation dominated by the N-fixing *Alnus viridis* (S8). Where the alpine prairie was enriched in dwarf Ericaceae, such as *Loiseleuria procumbens* and *Vaccinium uliginosum* subsp. *gaulterioides* (A5), the C/N ratio was higher. This ratio is characteristic of slowly decomposing organic matter, rich in fulvic and low-molecular weight acids, and is associated to low pH values. Indeed the fastest and strongest pH decrease was observed under subalpine forest/shrub vegetation, where values below 5 characterized the surface mineral horizon already after 60 years since deglaciation. Under grazed grassland and alpine prairie the pH reached comparable values after 130 and 260 years, respectively. Since the leaching of organic acids derived from organic matter degradation is the main acidifying factor in proglacial soils (Bernasconi et al. 2011), the slower pH decrease under alpine vegetation could be attributed to the lower biomass production and the stronger disturbances characterizing alpine habitats. However, a weak acidification was visible also in the youngest soils, extremely poor in organic matter, thanks to incipient mineral weathering and leaching caused by rain and snowmelt.

4.3 Evidences of podzolization processes

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

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

$$\text{ISratio} = 0.676 \cdot \text{age}^{0.185} \qquad (5).$$

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

Observing the IS and Fed depth trends (Figure 5 and Figure 6), the cheluviation seems to have actively translocated Fe and Al oxi-hydroxides into subsurface BC and CB horizons. The IS, in subsurface horizons of soils under subalpine forest-shrubs (*Rhododendron* and *Larch)*, increased according to the chronofunction (Figure 8b):

$IS=0.045+0.003*age^{0.8}$ (6)

therefore, the time needed was ca. 530 years.

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

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

Similarly to other chronosequences (e.g., Burt and Alexander 1996, Egli et al. 2001), the slower release and mobility of Al from the parent material, compared with Fe, could be related to the presence of easily weatherable Fe-rich minerals included in the small amounts of mafic rocks in the parent material. The sialic minerals weathered more slowly in the considered soils, as demonstrated by the Alo/Feo ratio increase with time, both in surface (not shown) and in subsurface horizons (Table 7). Conversely, in more acidic parent materials, the onset of the podzolization process tended to remove Al faster than Fe from the young and weakly developed E horizons (Mokma et al., 2004).

The E horizon formation in alpine soils has sometimes been interpreted as a stagnogley feature, caused by seasonal waterlogging and alternation of reductive and oxidative conditions, common in high altitude soils thanks to the abundant snow cover and high water input during the snowmelt (e.g. Gensac, 1990). However, the E horizon formation under subalpine vegetation in the Lys proglacial area cannot be ascribed specifically to seasonal waterlogging, as stronger seasonal reductive conditions should be met above timberline, thanks to deeper winter and spring snow covers. Here, E and CBs/Bs horizons were never found, and Fe/Al redistribution to subsurface horizons normally was not observed. Moreover the coarse texture of the soil material should facilitate drainage, decreasing the waterlogging potential in the topsoil.

The Feo/Fed ratio, called "activity ratio" and considered as a measure of the crystallinity of pedogenic Fe oxides (Burt and Alexander, 1996), did not show any significant time trends but had significantly lower values under subalpine forest-shrub vegetation than under grassland (Figure 9), particularly in subsurface horizons. Alternation of reductive and oxidative conditions are known to favor the crystallization of pedogenic Fe-oxi-hydroxides; however, the lower values found under subalpine vegetation, particularly in subsurface BC and Bs horizons, are not explainable according to surface seasonal redox conditions, for the same reason explained above. Nor the higher crystallinity found in soils under forest vegetation, compared to the soils developed under even more humid climates on the Alps (Dümig et al., 2011), could be easily explainable. Low Feo/Fed values, however, are typical of many well developed podzolic soils in the Valle d'Aosta Region (e.g. D'Amico et al., 2008).

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

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

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

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

5. Conclusions

In this study we characterize the main pedogenetic processes occurring in recently deglaciated areas under different vegetation covers. We furthermore provide evidence of fast rates of podzolization under a subalpine larch-Rhododendron forest, previously undocumented for the Alps: bleached E horizons are visible in 60 years old soils, and diagnostic albic horizons appear in 120 years. In this time span the cheluviation of organo-metal compounds in the underlying BC horizons was not sufficient to create diagnostic spodic features, but 300-500 years seemed enough for the formation of a diagnostic spodic Bs horizon and, consequently, of a "real" Podzol.

We suggest this relatively fast rate of podzolization be due to the specific plant community succession rather than to the climatic conditions in the study area, characterized by cold temperatures and significant but not exceptional precipitation amounts. The appearance of a larch-Rhododendron forest could significantly influence the soil organic matter characteristics, driving the soil development and the onset of soil forming processes in this proglacial area.

Acknowledgments

This study was accomplished thanks to the agreement with the "Centro Funzionale" of the Aosta Valley Region within the project "Optimization of the regional alert system regarding hydrogeological hazard, towards citizen protection: soil vulnerability map, 1:100000", and thanks to the Italian MIUR Project (PRIN 2010-11): "Response of morphoclimatic system dynamics to global changes and related geomorphological hazards" (national coordinator C. Baroni). We also thank three anonymous reviewers for the great suggestions on how to improve the quality of the paper.

References

Abakumov, E., Trubetskoj, O., Demin, D., Celi, L., Cerli, C., Trubetzksaya, O., 2010. Humic acid characteristics in a Podzol soil chronosequence. Chem. Ecol. 26:S2, 59-66

Alexander, E.B., and Burt, R., 1996. Soil development on moraines of the Mendenhall Glacier, southeast Alaska. 1. The moraines and soil morphology. Geoderma 72, 1-17

Anderson, S.P., Drever, J.I., Frost, C.D., Holden P,. 2000. Chemical weathering in the foreland of a retreating glacier. Geochim. Cosmochim. Acta 64, 1173-1189.

Andreis, C., Caccianiga, M., Cerabolini, B., 2001. Vegetation and environmental factors during primary succession on glacier forelands: some outlines from the Italian Alps. Plant Biosystems 135(3), 295-310.

Bernasconi, S.M., Bauder, A., Bourdon, A., Brunner, I., Bünemann, E., Christl, I., Derungs, N., Edwards, P., Farinotti, D., Frey, B., Frossard, E., Furrer, G., Gierga, M., Göransson, H., Gülland, K., Hagedorn, F., Hajdas, I., Hindshaw, R., Ivy-Ochs, S., Jansa, J., Kiczka, M., Kretschmar, R., Lemarchand, E., Luster, J., Magnusson, J., Mitchell, E.A.D., Venterink, H., O., Plötze, M., Reynolds, B., Smittenberg, R., H., Stähli, M., Tamburini F., Tipper, E.T., Wacker, L., Welc, M., Wiederhold, J.G., Zeyer, J., Zimmermann, S., and Zumsteg, A., 2011. Chemical and biological gradients along the Damma Glacier soil chronosequence, Switzerland. Vadose Zone J. 10, 867-883.

Bernier, N., Gillet, F., 2012. Structural relationships among vegetation, soil fauna and humus form in a subalpine forest ecosystem: a Hierarchical Multiple Factor Analysis (HMFA). Pedobiologia 55, 321-334

Boettcher, S.E., Kalisz, P.J., 1990. Single-Tree Influence on Soil Properties in the Mountains of Eastern Kentucky. Ecology 71 (4), 1365-1372

Boggs, K., Klein, S.C., Grunblatt, J., Boucher, T., Koltun, B., Sturdy, M., Streveler, G.P., 2010. Alpine and subalpine vegetation chronosequences following deglaciation in coastal Alaska. Arctic, Antarctic and Alpine Res. 42(4), 385-395.

Braun-Blanquet, J., 1948. Übersicht der pflanzengesellschaften rhaetiens (IV). Vegetatio 2(1), 20-37.

Burga, C., 1999. Vegetation development on the glacier forefield Morteratsch (Switzerland). Applied Veg. Sci. 2, 17-24.

Burga, C.A., Krüsi, B., Egli, M., Wernli, M., Elsener, S., Ziefle, M., Fischer, T., Mavris, C., 2010. Plant succession and soil development on the foreland of the Morteratsch glacier (Pontresina, Switzerland): straight forward or chaotic? Flora 205, 561-576.

Burt, R, and Alexander, E. B., 1996. Soil development on moraines of Mendenhall Glacier, southeast Alaska. 2. Chemical transformations and soil micromorphology. Geoderma 72,19-36.

Caccianiga, M., Luzzaro, A., Pierce, S., Ceriani, R. M, Cerabolini, B., 2006.The functional basis of a primary succession resolved by CSR classification. Oikos 112, 10-20

Dümig, A., Smittenberg, R., Kögel-Knaber, I., 2011. Concurrent evolution of organic and mineral components after retreat of the Damma glacier, Switzerland. Geoderma 163, 83-94

Egli, M., Fitze, P., Mirabella, A., 2001. Weathering and evolution of soils formed on granitic, glacial deposits: results from chronosequences of Swiss alpine environments. Catena 45, 19-47.

Egli, M., Mirabella, A, Fitze, P., 2003. Formation rates of smectite derived from two Holocene chronosequences in the Swiss Alps. Geoderma 117, 81-98

Egli, M., Wernli, M., Kneisel, C., Haeberli, W., 2006. Melting glaciers and soil development in the proglacial area Morteratsch (Swiss Alps): I. Soil type chronosequences. Arctic, Antarctic and Alpine Res. 38(4), 499-509

FAO, 2006. Guidelines for Soil Description. 4th ed. FAO, Rome, 97 pp.

Föllmi, K.B., Arn, K., Hosein, R., Adatte, T., Steinmann, 2009. Biogeochemical weathering in sedimenraty chronosequences of the Rhone and Oberaar glacies(Swiss Alps): rates and mechanisms of biotite weathering. Geoderma 151, 270-281

Forman, S.L., Miller, G.H., 1984. Time-dependent soil morphologies and pedogenic processes on raised beaches, Bröggerhalvöya,Spitsbergen, Svalbard archipelago. Arctic and Alpine Res. 16, 4, 381-394

He, L., Tang, Y., 2008. Soil development along primary succession sequences on moraines of Hailuogou Glacier, Gongga Mountain, Sichuan, China. Catena 72, 259-269

Hoffland, E., Kuyper, T.W., Wallander, H., Passard, C., Gorbushina, A.A., Haselwandter, K., Holmström, S., Landeweert, R., Lundström, U.S., Rosling, A., Sen, R., Smits, M.m., van Hees, P.A.W., van Breemen, N., 2004. The role of fungi in weathering. Frontiers in Ecology and the Environment 2, 5, 258-264

Hodkinson, I.D., Coulson, S.J., Webb, N.R., 2003. Community assembly along proglacial chronosequences in the high Arctic: vegetation and soil development in north-west Svalbard. J. Ecol., 91, 651-663

Hothorn, T., Bretz, F., and Westfall, P. (2008). Simultaneous Inference in General Parametric Models. Biometr. J. 50(3), 346—363.

Huggett, R.J., 1998. Soil chronosequences, soil development, and soil evolution: a critical review. Catena 32, 155-172

IUSS Working Group WRB, 2006. World reference base for soil resources 2006. World Soil Resources Reports No. 103. FAO, Rome.

Ivy-Ochs, S., Kerschner, H., Maisch, M., Christl, M., Kubik, P.W., and Schluechter, C., 2009. Latest Pleistocene and Holocene glacier variations in the European Alps: Quat. Sci. Rev. 28, 2137–2149

Jeanroy, E., Guillet, B., Ortiz, R., 1984. Applications pédogénétiques de l'étude des forms du fer par les réactifs d'extraction: cas des sols brunifiés et podzolisés sur roches cristallines. Sci. Sol 3, 199–211.

Jenny, H., 1958. Role of the plant factor in the pedogenic function. Ecology 39 (1), 5-46

Joerin, U.E., Stocker, T.F., Schlüchter, C. 2006. Multicentury glacier fluctuations in the Swiss Alps during the Holocene. The Holocene 16 (5), 697-704

Koele, N., Storch, F., Hildebrand, E.E., 2011. The coarse-soil fraction is the main living space of fungal hyphae in the BhBs horizon of a Podzol. J. Plant Nutr. Soil Sci. 174, 750–753

Landeweert, R., Hoffland, E., Finlay, R.D., Kuyper, T.W., van Breemen, N, 2001. Linking plants to rocks: ectomycorrhizal fungi mobilize nutrients from minerals. Trends in Ecol. & Evol. 16, 5, 248-254

Landolt, E., (1977): Ökologische Zeigerwerte zur Schweizer Flora. Veröffentlichungen des Geobotanischen Institutes der Eidgenössischen Technischen Hochschule, Stiftung Rübel, Zürich 64: 1–208.

Lichter, J., 1998. Rates of weathering and chemical depletion in soils across a chronosequence of Lake Michigan sand. Geoderma 85, 255-282

Lundström, U.S., van Breemen, N., Bain, D.C., van Hees, P.A.W., Giesler, R., Gustafsson, J.P., Ilvesniemi, H., Karltun, E., Melkerud, P.A., Olsson, M., Riise, G., Wahlberg, O., Bergelin, A., Bishop, K., Finlay., R., Jongmans, A.G., Magnusson, T., Mannerkoski, H., Nordgren, A., Nyberg, l., Starr, M., Tau Strandm L., 2000. Advances in understanding the podzolization process resulting from a multidisciplinary study of three coniferous forest soils in the Nordic Countries. Geoderma 94, 335-353

Mahaney, W.C., Kalm, V., Milner M.W., Hancock, R.G.V., 2009. A soil chronosequence in late glacial and neoglacial moraines, Humboldt Glacier, northwestern Venezuela Andes. Geomorphology 109, 236-245.

Mattirolo, E., Novarese, V., Franchi, S., Stella, A., 1951. Carta Geologica d'Italia 1:100000, foglio 29. Istituto Geografico Militare (Firenze, Italy)

Mellor, A., 1987. A pedogenic investigation of some soil chronosequences on neoglacial moraine ridges, Southern Norway: examination of soil chemical data using principal component analysis. Catena 14, 369-381

Mercalli L., 2003. Atlante climatico della Valle d'Aosta. Società Meteorologica Italiana (Ed), Torino, p. 405.

Michalet, R., Gandoy, C., Cadel, G., Girard, G., Grossi, J., Joud, D., Pache, G., 2001. Modes de fonctionnement d'humus des forêts sempervirentes des Alpes internes françaises. Humus functioning types in evergreen coniferous forests of the French Inner Alps. 324 (1), 59-70

Ministero delle Politiche Agricole e Forestali, 2000. Metodi di Analisi Chimica dei Suoli. In: Violante P, Sequi P (eds), Collana di Metodi Analitici per l'Agricoltura. Franco Angeli (Ed.), Milano, pp 1–474

Mokma, D.L., Yli-Halla, M., Lindqvist, K., 2004. Podzol formation in sandy soils of Finland. Geoderma 120, 259-272

Monterin U., 1932. Le variazioni secolari del clima del Gran S. Bernardo: 1918-1931 e le oscillazioni del ghiacciaio del Lys al Monte Rosa: 1789-1931. Boll. Com. Glac. It., ser. I, 12. 59-189

Ochs M., 1996. Influence of humified and non-humified natural organic compounds on mineral dissolution. Chem. Geol. 132, 119-124

Pelfini, M., Belloni, S., Rossi, G., Struma, G., 1997. Geogr. Fis. Din. Quat. 20, 329-338

Pignatti, S., 1992. Flora d'Italia, vol 1–3. Edagricole, Bologna

Pornon, A., Doche, B., 1996. Age Structure and Dynamics of Rhododendron ferrugineum L. populations in theNorthwestern French Alps. J. Veg. Sci. 7(2), 265-272

Raffl, C., Mallaun, M., Mayer, R., Erschbamer, B., 2006. Vegetation succession pattern and diversity changes in a glacier valley, central Alps, Austria. Arctic, Antarctic and Alpine Res. 38(3), 421-428.

Righi, D., Huber, K., Keller, C., 1999. Clay formation and Podzol development from postglacial moraines in Switzerland. Clay Min. 34, 319-322

Sauer, D., Sponagel, H., Sommer, M., Giani, L., Jahn, R., Stahr, K., 2007. Podzol: soil of the year 2007; a review of its genesis, occurrence, and functions. J. Plant Nutr. Soil Sci. 170, 581-597

Schaetzl, R., Anderson, S., 2005, Soils. Genesis and geomorphology. Cambridge University Press (Ed), Cambridge (UK)

Soil Survey Staff, 2010. Keys to Soil Taxonomy. United States Department of Agriculture, Natural Resources Conservation Service, Eleventh Edition, 2010

Strada, E., 1988. Le variazioni del ghiacciaio del Lys dalla "Piccola Glaciazione" ai nostri giorni. Natura bresciana, Ann. Mus. Civ. Sc. Nat. 24: 275-188

Tonkin, P.J., Basher, L.R., 2001. Soil chronosequences in subalpine superhumid Cropp Basin, western Southern Alps, New Zealand. New Zealand J. Geol. Geophys. 44 (1), 37-45

Treter, U., Ramsbeck-Ullmann, M., Böhmer, H.J., Bösche, H., 2002. Vegetationsdynamic im vorfeld des Lys-gletschers (Valle di Gressoney/Region Aosta/Italien). Erdkunde 56, 253-267

Ugolini, F.C. 1966. Part 3. Soils. In: A. Mirskey, (Ed,). Soil development and ecological succession in a deglaciated area of Muir Inlet, southeast Alaska. Institute of Polar Studies report Number 20, Ohio State University, Columbus, Ohio, USA.

Ugolini FC, Corti G, Certini G. (2006). Pedogenesis in the sorted patterned ground of Devon Plateau,Devon Island, Nunavut, Canada. Geoderma 136, 87-106

Van Breemen, N., Lundström, U.S., Jongmans, A.G., 2000. Do plants drive podzolization via rock-eating mycorrhizal fungi? Geoderma 94. 163–171

Zanella A, Jabiol B, Ponge JF, Sartori G, De Waal R, Van Delft B, Graefe U, Cools N, Katzensteiner K, Hager H, Englisch M, Brethes A, Broll G, Gobatl JM, Brun JJ, Milbert G, Kolb E, Wolf U, Frizzera L, Galvan P, Kolli R, Baritz R, Kemmerse R, Vacca A, Serra G, Banas D, Garlato A, Chersich S, Klimo E, Langohr R (2011a) European Humus Forms Reference Base.

Zech, W., Wilcke, B.M., 1977. Vorlauge ergebnisse einer Bodenchronosequenzstudie im Zillertal. Mitteilungen der Deutschen Bodenkundlichen Geselschaft 25, 571-586

Figure captions

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

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

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

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

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

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

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

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

Fig. 9: activity ratio in surface and subsurface horizons of grassland and forest soils (p-value < 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						
Cerastium uniflorum	1	3	0	0	0	0
Epilobium fleischeri	1	5	2	0	0	0
Linaria alpina	1	1				
Oxyria digyna	1	0	0	0	0	0
Saxifraga oppositifolia	0	0	1	0	0	0
Trisetum distichophyllum	1	3	0	0	0	0
Mid-successional						
Achillea nana	0	5	0	5	0	0
Agrostis schraderiana	1	0	0	0	0	0
Anthyllis vulneraria subsp *valesiaca*	0	2	2	0	0	0
Aster alpinus	0	1	1	0	0	0
Campanula cochleariifolia	0	0	0	1	0	0
Carex norvegica	0	0	1	0	0	0
Carex ornithopodioides	0	0	1	0	0	0
Erigeron alpinus	0	1	1	0	0	0
Gentiana nivalis	0	0	1	0	0	0
Luzula spicata	0	0	2	3	0	0
Myosotis alpestris	0	0	10	10	0	0
Saxifraga exarata	1	0	2	0	0	0
Saxifraga paniculata	0	0	2	0	0	0
Silene excapa	1	0	5	1	0	0
Silene rupestris	0	0	0	1	0	0
Trifolium pallescens	0	30	10	5	1	0
Late successional						
Antennaria dioica	0	0	0	0	1	0
Anthoxanthum alpinum	0	0	1	0	0	0
Bellis perennis	0	0	0	0	1	0
Botrychium lunaria	0	0	0	1	0	0
Carex curvula	0	0	20	5	5	0
Carex sempervirens	0	0	0	0	0	5
Coeloglossum viride	0	0	0	0	1	0
Euphrasia rohoskoviana	0	0	3	0	0	0
Festuca halleri	0	1	3	5	2	2

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

Sempervivum montanum	0	3	3	2	0	0
Veronica bellidioides	0	0	1	0	0	0
Veronica fruticans	0	10	0	0	0	0

Table 3: Subalpine 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-mid successional species, grassland and ubiquitous species and subalpine forest (climax) species was based on Burga et al. 2010 and on Pignatti (1992).

	P1	SG2	S3	S4	SG5	S6	SG7	S8	S9	S10
Early-mid successional										
Achillea nana	0	1	2	0	1	0	0	0	0	0
Cerastium uniflorum	1	0	0	0	0	0	0	0	0	0
Epilobium fleischeri	1	1	1	0	0	0	0	0	0	0
Linaria alpina	1	0	0	0	0	0	0	0	0	0
Oxyria digyna	1	0	0	0	0	0	0	0	0	0
Poa nemoralis	0	10	0	0	2	0	0	0	0	0
Rumex scutatus	0	0	1	0	0	0	0	0	0	0
Salix appendiculata	0	20	5	0	10	0	0	0	0	0
Salix helvetica	0	5	0	0	10	0	0	0	0	0
Saxifraga exarata	1	0	0	0	0	0	0	0	0	0
Silene excapa	1	0	0	0	0	0	0	0	0	0
Trisetum distichophyllum	1	0	1	0	0	0	0	0	0	0
Grassland and ubiquitous species (selection)										
Agrostis schraderiana	1	5	5	0	0	0	10	1	0	0
Anthoxanthum alpinum	0	1	0	0	1	0	0	0	10	0
Anthyllis vulneraria subsp *valesiaca*	0	0	2	0	0	0	0	0	5	0
Astragalus penduliflorus	0	0	0	0	0	0	0	0	5	0
Calluna vulgaris	0	0	0	0	0	0	0	0	5	0
Campanula barbata	0	0	0	0	0	0	0	0	1	0
Carduus defloratus	0	0	0	0	0	0	0	0	1	0
Carex sempervirens	0	0	2	0	0	0	0	0	2	0
Carlina acaulis	0	0	1	0	0	0	0	0	0	0
Cerastium arvense	0	3	0	0	0	0	0	0	1	0

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

Astrantia minor	0	0	0	0	0	0	0	2	0	0
Avenella flexuosa	0	15	0	0	0	10	50	5	0	5
Calamagrostis villosa	0	5	0	0	3	0	0	0	0	10
Dactylorhiza maculata subsp fuchsii	0	0	0	0	0	0	0	1	0	0
Hieracium murorum aggr.	0	0	0	0	0	0	0	3	0	0
Homogyne alpina	0	0	0	0	0	0	0	2	0	5
Juniperus communis	0	0	0	0	0	10	0	0	0	5
Larix decidua	0	40	30	0	60	50	80	80	0	60
Orthilia secunda	0	3	0	0	0	0	0	0	0	0
Peucedanum ostruthium	0	0	0	0	0	0	0	5	0	0
Phyteuma betonicifolium	0	1	0	0	0	0	0	0	1	0
Rhododendron ferrugineum	0	0	30	0	80	60	50	60	0	70
Vaccinium myrtillus	0	0	0	0	0	40	50	30	0	30
Vaccinium uliginosum subsp. gaulterioides	0	0	20	0	0	0	0	0	0	20
Vaccinium vitis-idaea	0	0	0	0	2	20	0	0	0	20
Viola biflora	0	0	0	0	0	0	0	1	0	1

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
		C2	16-35+	2.5Y5/3	81.5	1.0	AB	LO	60	2
A3	c. 1950	OL	0-0.5							
		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
A4	1821	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
A5	c. 1755	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	
		C	40-50+	2.5Y5/4	94.2	0.5	AB	LO	30	
A6	c. 11000 BP	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
	C2	5.6	0.17	0.80	0.18	1.40	0.57
A3	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
	C	6.5	0.07	1.73	0.21	2.15	0.77
A4	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
	C	5.1	0.31	1.30	0.40	2.80	0.46
A5	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

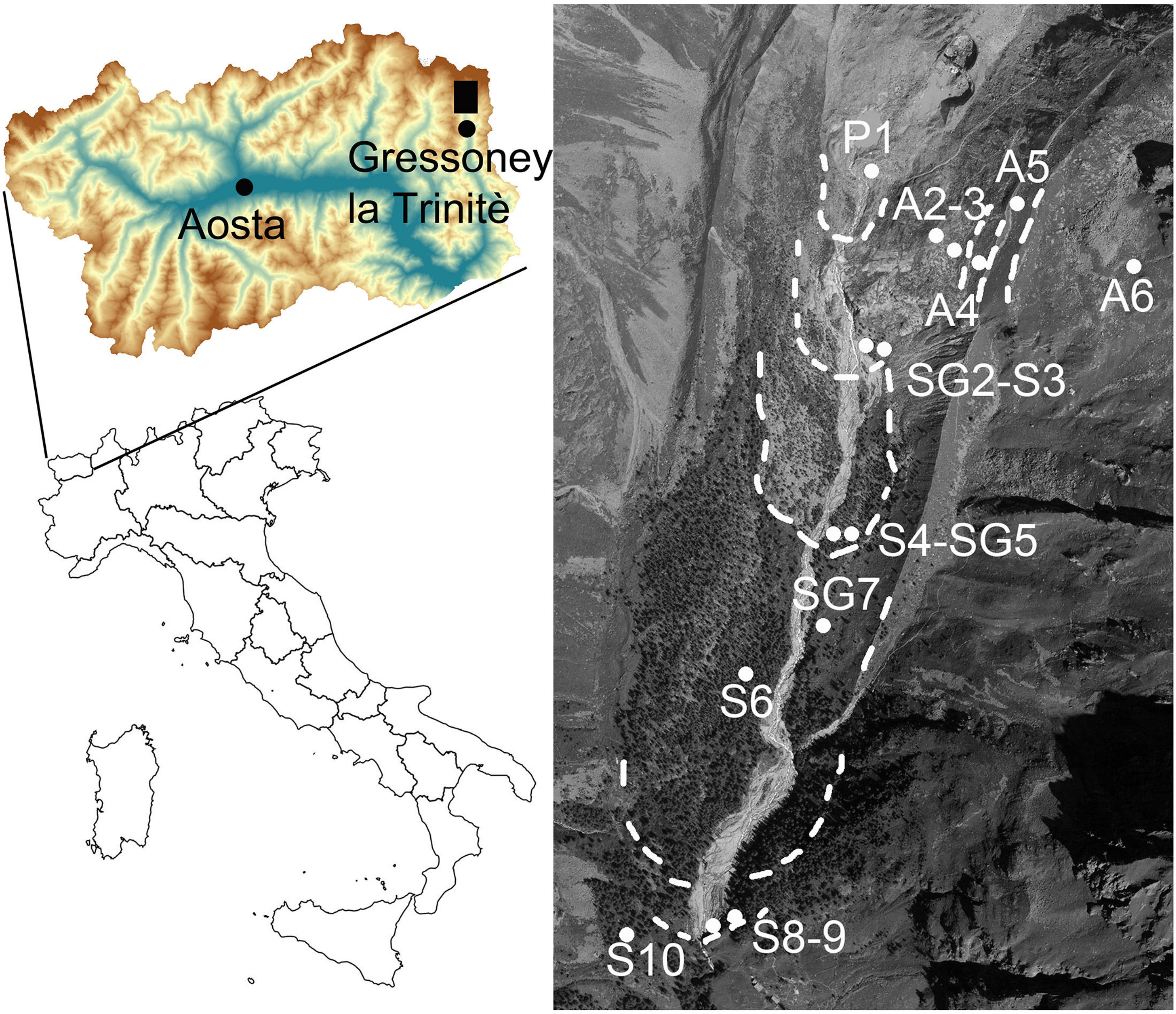
Tab. 6: Macromorphological properties of the soils in the subalpine chronosequence. Structure: GR = granular; PL = platy; PS = subangular polyhedral; MA = massive; RS = rock structure; AB = absent; M = matted (O horizons). The number indicates the size class of the soil structure: 1 = very fine, 2 = 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).

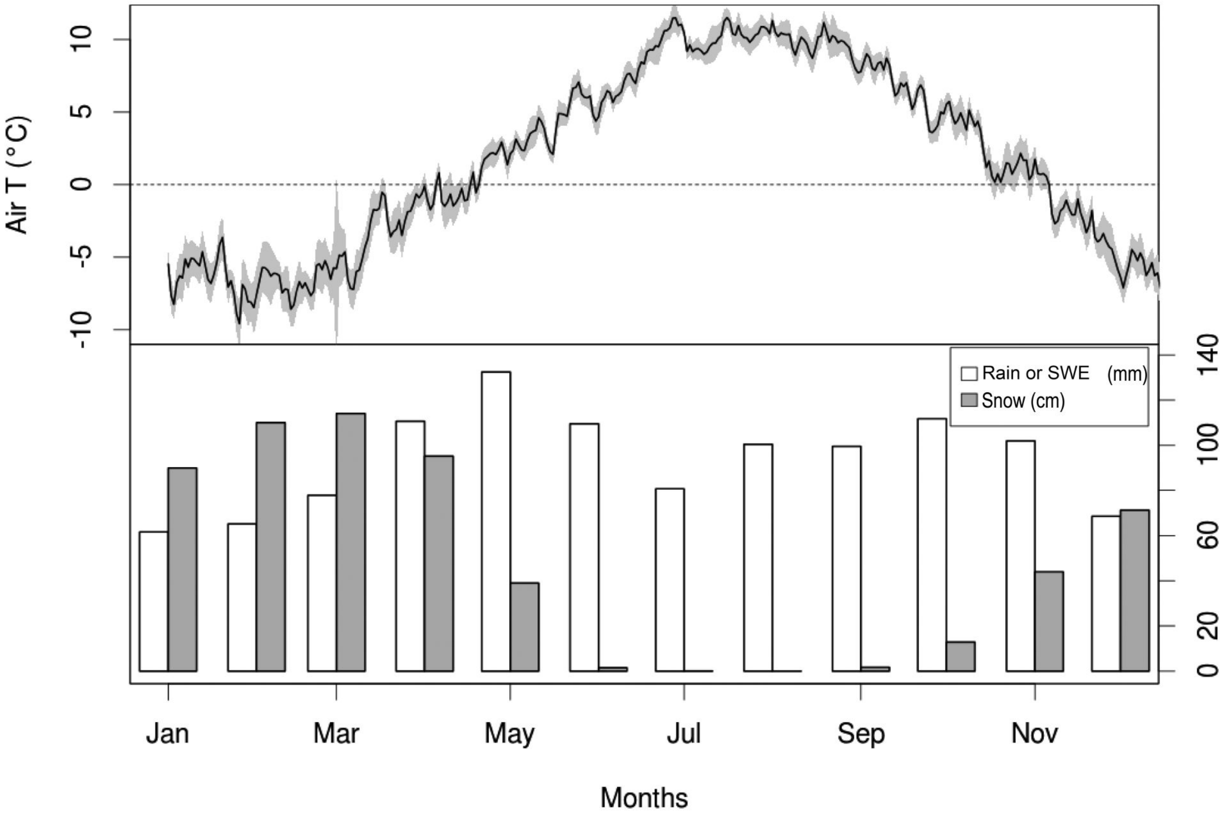
Profile	Year of deposition	Hor.	Depth (cm)	Munsell Color (moist)	Sand (%)	Clay (%)	Structure (or cementation)	Consistence (moist)	Rock fragments (volume, %)	Silt caps
P1	2004	C1	0-8	5Y5.5/3	69.1	2.1	PL3/AB	LO	60	
		C2	8-47+	5Y6/1	80.3	0.8	PL3/AB	LO	60	1
SG2	c. 1950	OL	0-1							
		OH	1-3	7.5YR2/2						
		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	
		C	18-40+	2.5Y5/3	86.4	0.8	M	M	50	3
S3	c. 1950	OL	0-1							
		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	
S4	1921	OL	0-1							
		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	
SG5	1921	OF	0-4	10YR2/1			GR1	LO		
		A	4-17	10YR4/3	80.5	1.7	GR1	LO	60	
		C	17-30+	10YR5/3	81.7	0.8	AB	LO	80	1
S6	c. 1880	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	
		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	
SG7	c. 1880	OH	0-1							
		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
S8	1821	OL	0-2							

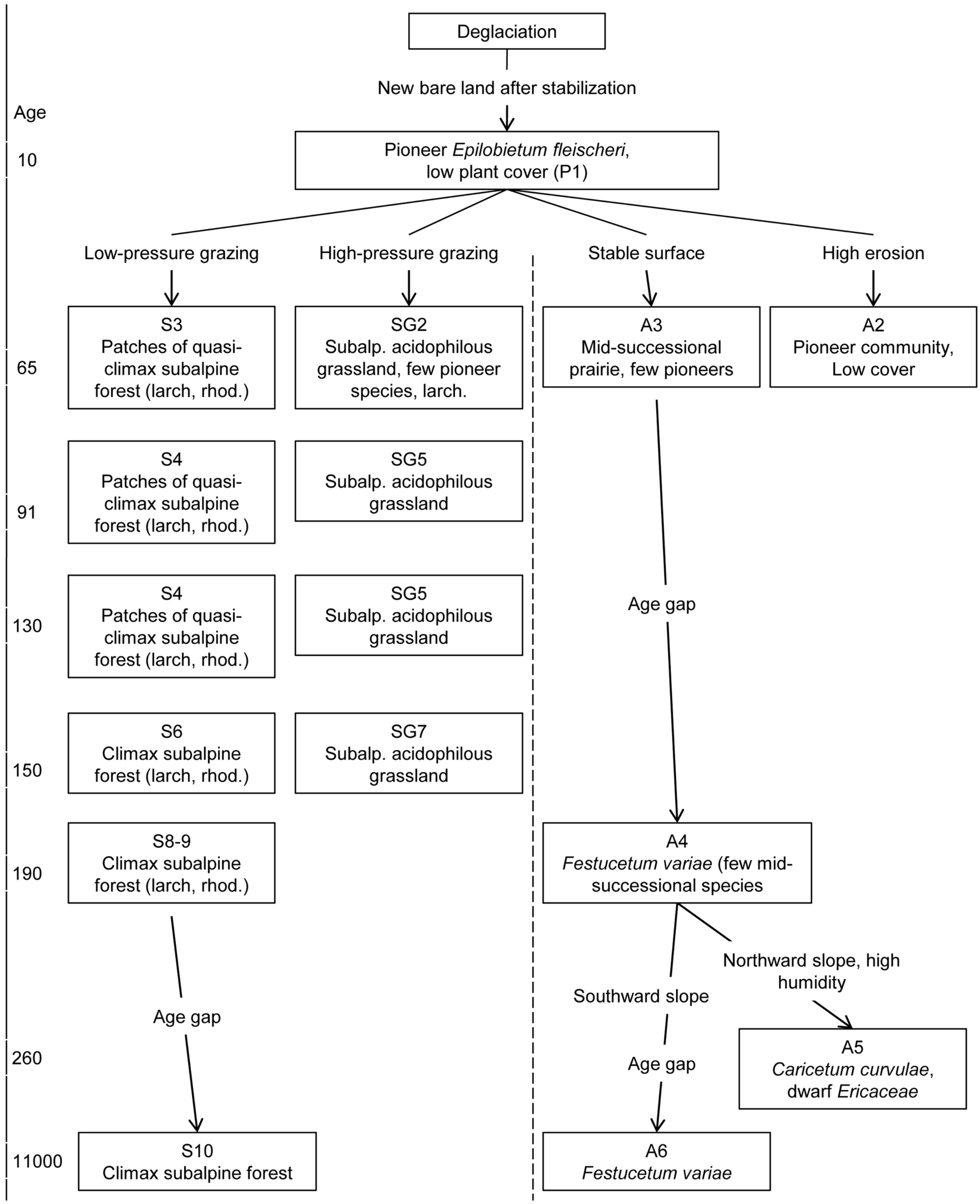
		OF	2-4	7.5YR2.5/2			M1	LO	
		E	4-7	10YR5.5/2, 10YR3/1	67.5	2.1	GR1	LO1	50
		BCs	7-21	10YR5/4	81.9	1.0	GR1	LO1	60
		C	21-42	10YR5/3	82.6	0.8	AB (PS1)	LO1	80
S9	1821	OL	0-1						
		OF	1-5	5YR2/2					
		OH	5-10/7	5YR2/2					
		AE	10/7-12	10YR4/3	68.2	1.9	GR0	M	30
		E	12-18	10YR5/2	66.5	2.1	GR1	M	40
		BCs	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

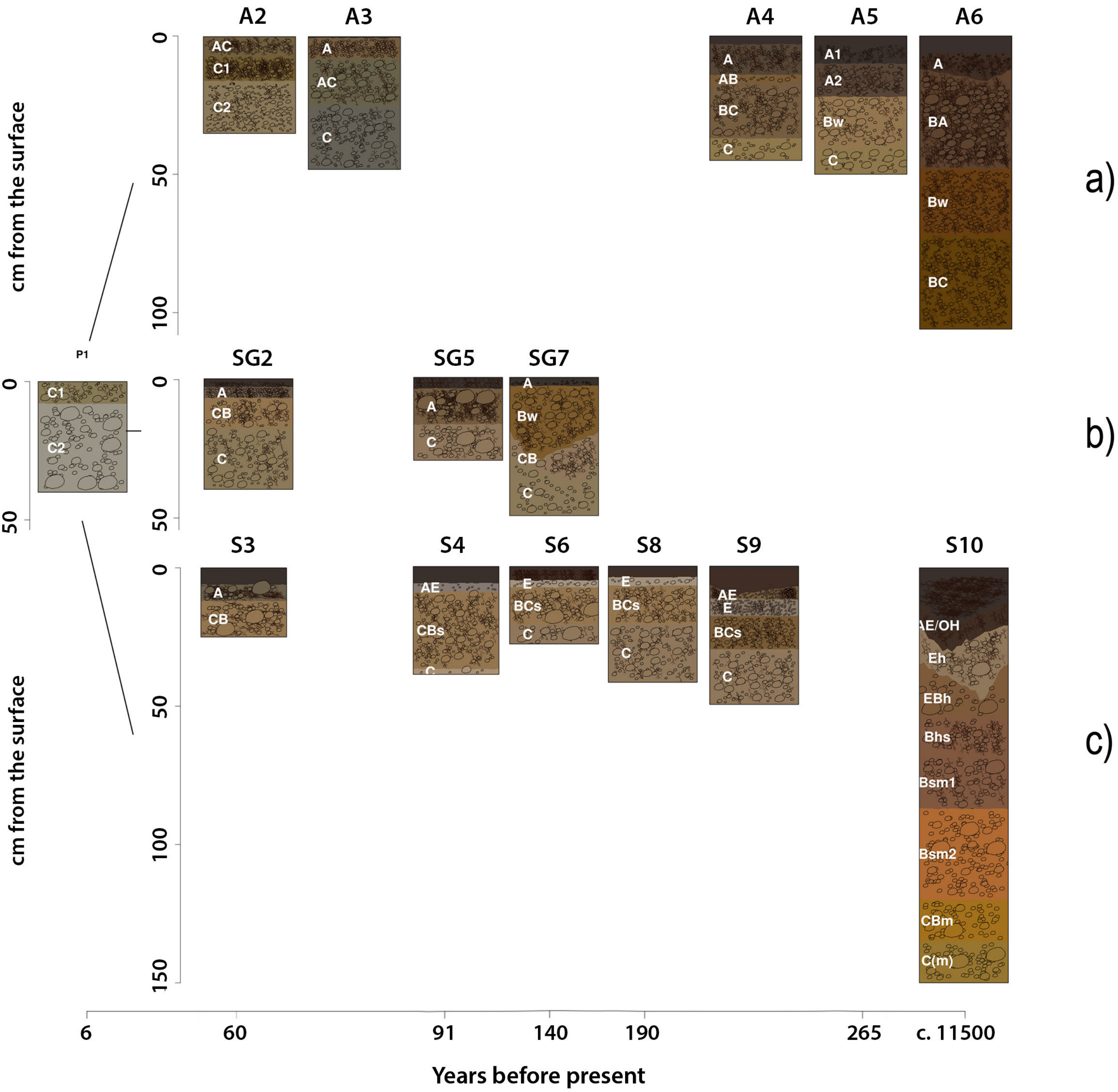
Tab. 7: Chemical properties of the subalpine soils along the Lys forefield chronosequences.

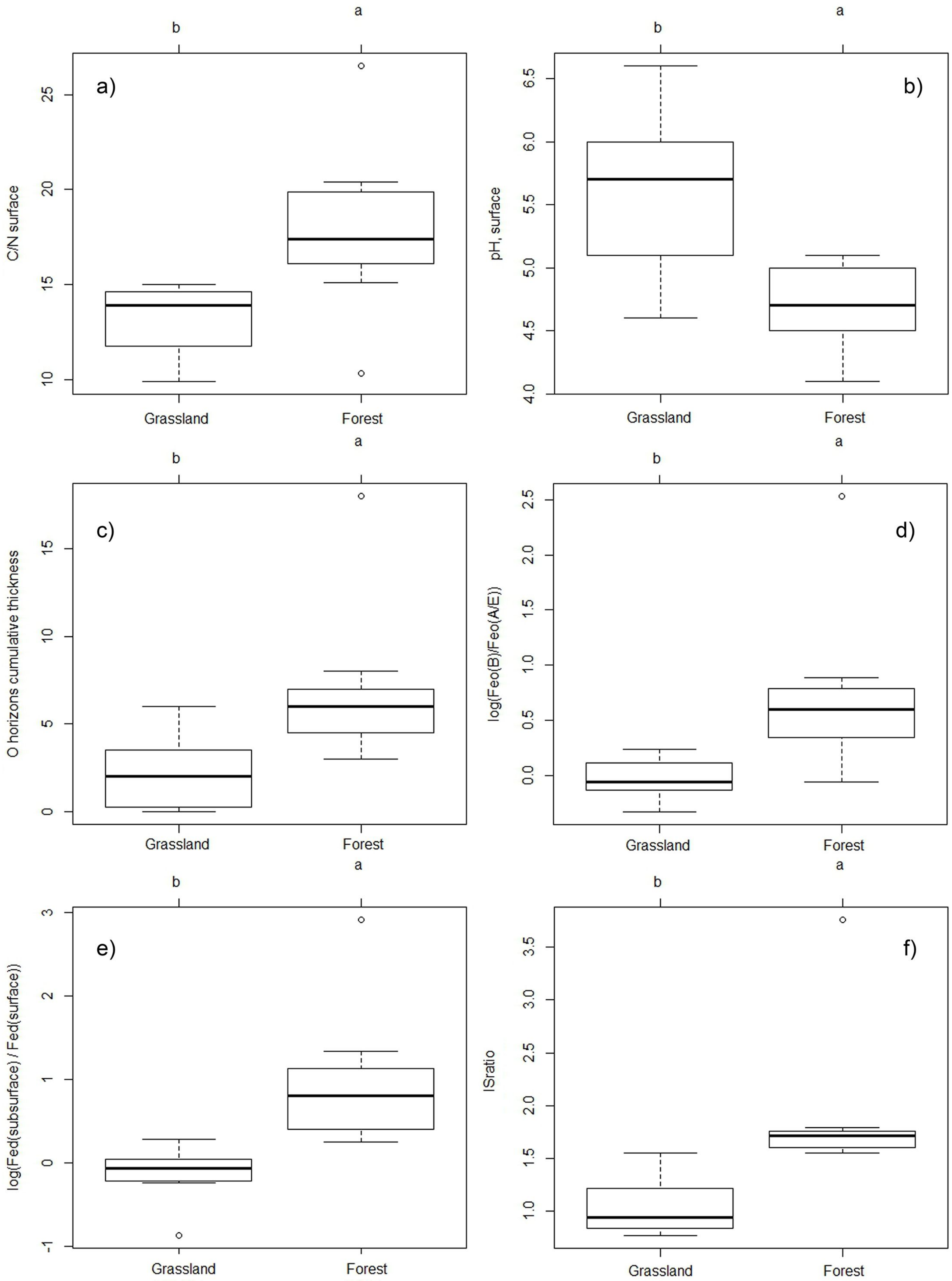
	Horizon	pH	C	Feo	Alo	Fed	Feo/Fed	Alo/Feo
			%	%	%	%		
P1	C1	6.4	0.04	0.09	0.20	0.19	0.47	0.22
	C2	6.6	0	0.08	0.18	0.15	0.53	0.25
SG2	OH	5.9	26.58					
	A	5.8	1.5	0.11	0.27	0.19	0.58	0.27
	CB	5.8	0.44	0.14	0.36	0.25	0.56	0.29
	C	5.9	0.15	0.10	0.31	0.21	0.48	0.30
S3	OL/OF	5.6	26.89					
	A	5	1.92	0.07	0.19	0.18	0.39	0.29
	CB	5.2	0.68	0.11	0.31	0.23	0.48	0.27
S4	OL/OF	5.1	41.2					
	OH	4.9	18.44					
	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
SG5	OF	5.8	14.32					
	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
S6	OF-OH	4.4	35.24					
	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
SG7	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
	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
S8	OL-OF	4.6	16.32					
	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


Gressoney
la Trinitè
Aosta
P1
A5
A2-3
A4
A6
SG2-S3
S4-SG5
SG7
S6
S8-9
S10

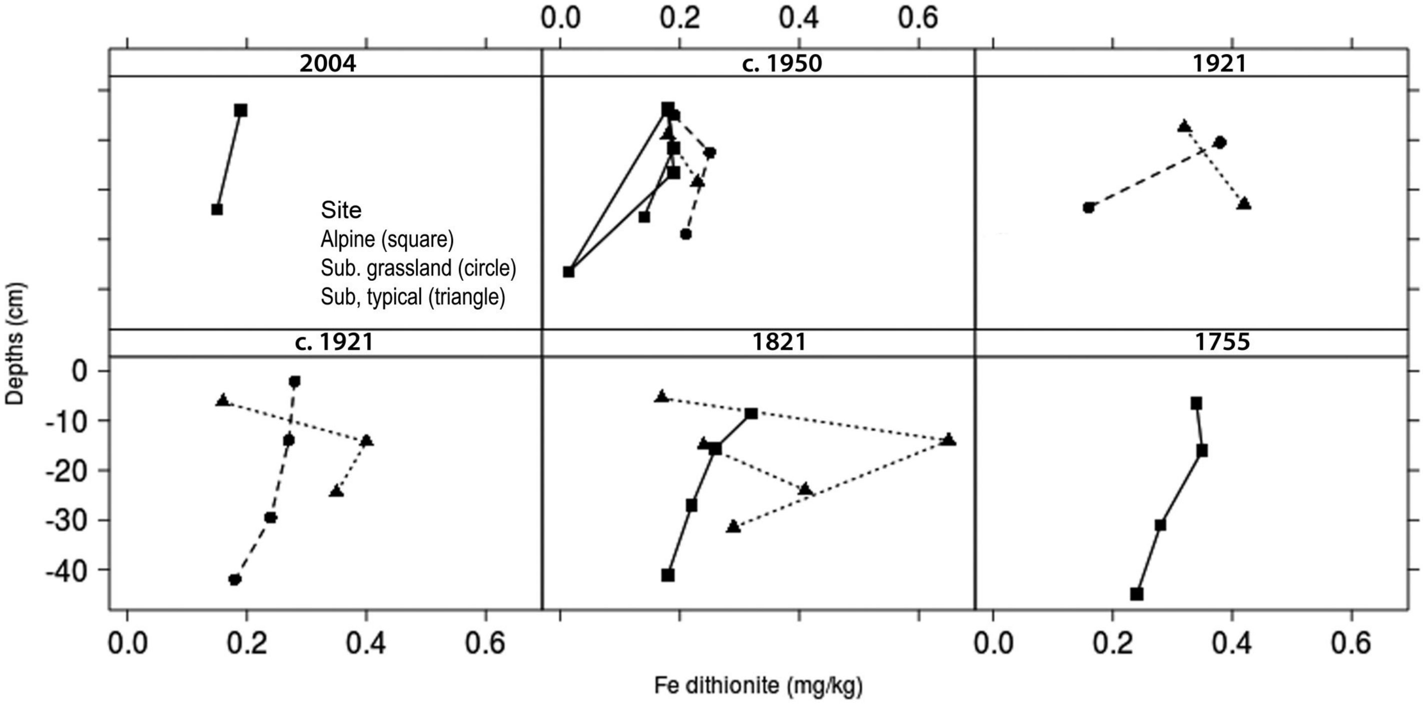

Air T (°C)
10
5
0
-5
-10
Rain or SWE (mm)
Snow (cm)
140
100
60
20
0
Jan
Mar
May
Jul
Sep
Nov
Months


Age
10
65
91
130
150
190
260
11000
Deglaciation
New bare land after stabilization
Pioneer *Epilobietum fleischeri*, low plant cover (P1)
Low-pressure grazing
High-pressure grazing
Stable surface
High erosion
S3 Patches of quasi-climax subalpine forest (larch, rhod.)
SG2 Subalp. acidophilous grassland, few pioneer species, larch.
A3 Mid-successional prairie, few pioneers
A2 Pioneer community, Low cover
S4 Patches of quasi-climax subalpine forest (larch, rhod.)
SG5 Subalp. acidophilous grassland
S4 Patches of quasi-climax subalpine forest (larch, rhod.)
SG5 Subalp. acidophilous grassland
Age gap
S6 Climax subalpine forest (larch, rhod.)
SG7 Subalp. acidophilous grassland
S8-9 Climax subalpine forest (larch, rhod.)
A4 *Festucetum variae* (few mid-successional species
Northward slope, high humidity
Southward slope
Age gap
Age gap
A5 *Caricetum curvulae*, dwarf *Ericaceae*
S10 Climax subalpine forest
A6 *Festucetum variae*

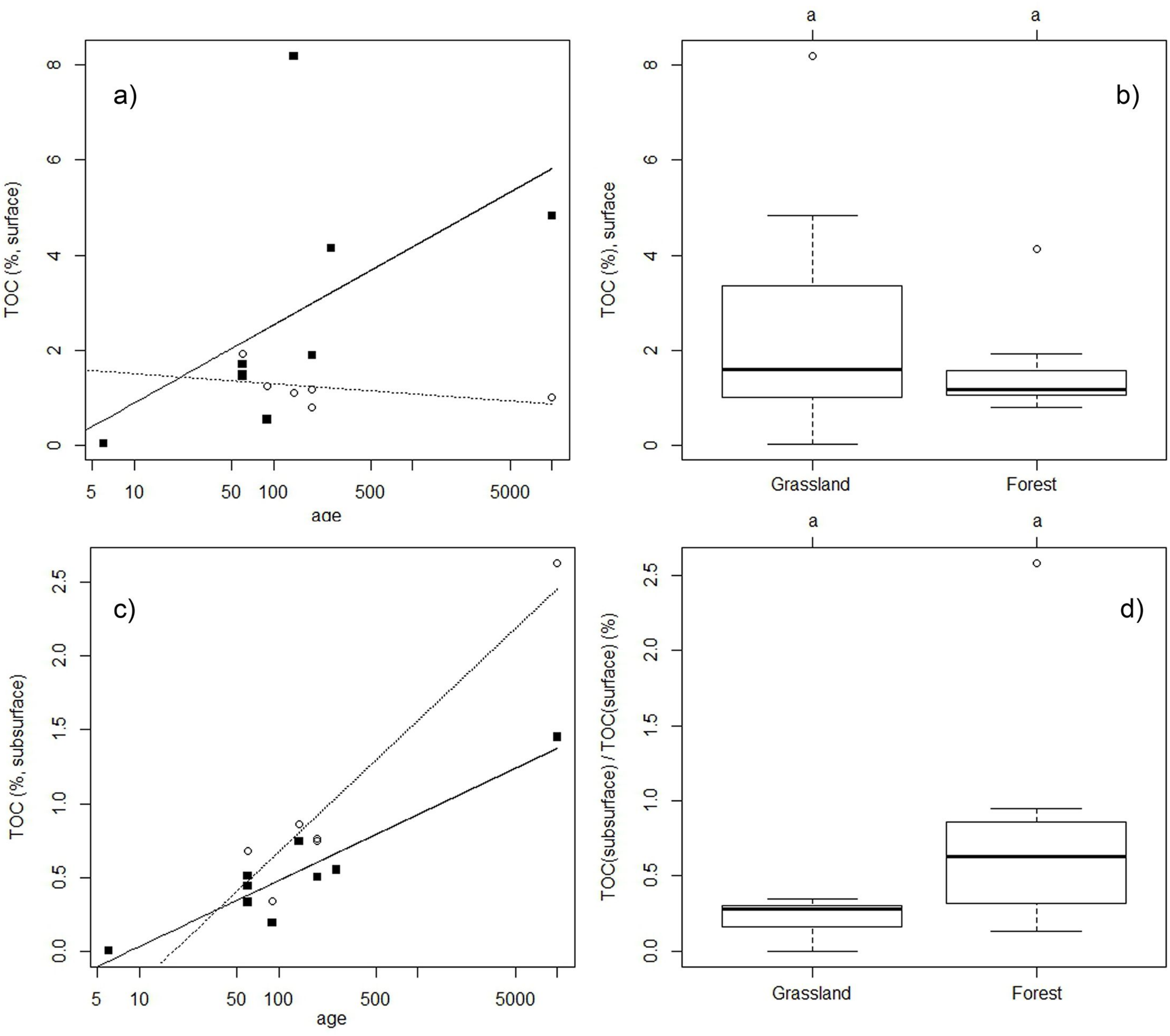

A2
A3
A4
A5
A6
AC
C1
C2
A
AC
C
A
AB
BC
C
A1
A2
Bw
C
A
BA
Bw
BC
a)
P1
C1
C2
SG2
A
CB
C
SG5
A
C
SG7
A
Bw
CB
C
b)
S3
A
CB
S4
AE
CBs
C
S6
E
BCs
C
S8
E
BCs
C
S9
AE
E
BCs
C
S10
AE/OH
Eh
EBh
Bhs
Bsm1
Bsm2
CBm
C(m)
c)
cm from the surface
cm from the surface
0
50
100
0
50
0
50
0
50
100
150
6
60
91
140
190
265
c. 11500
Years before present



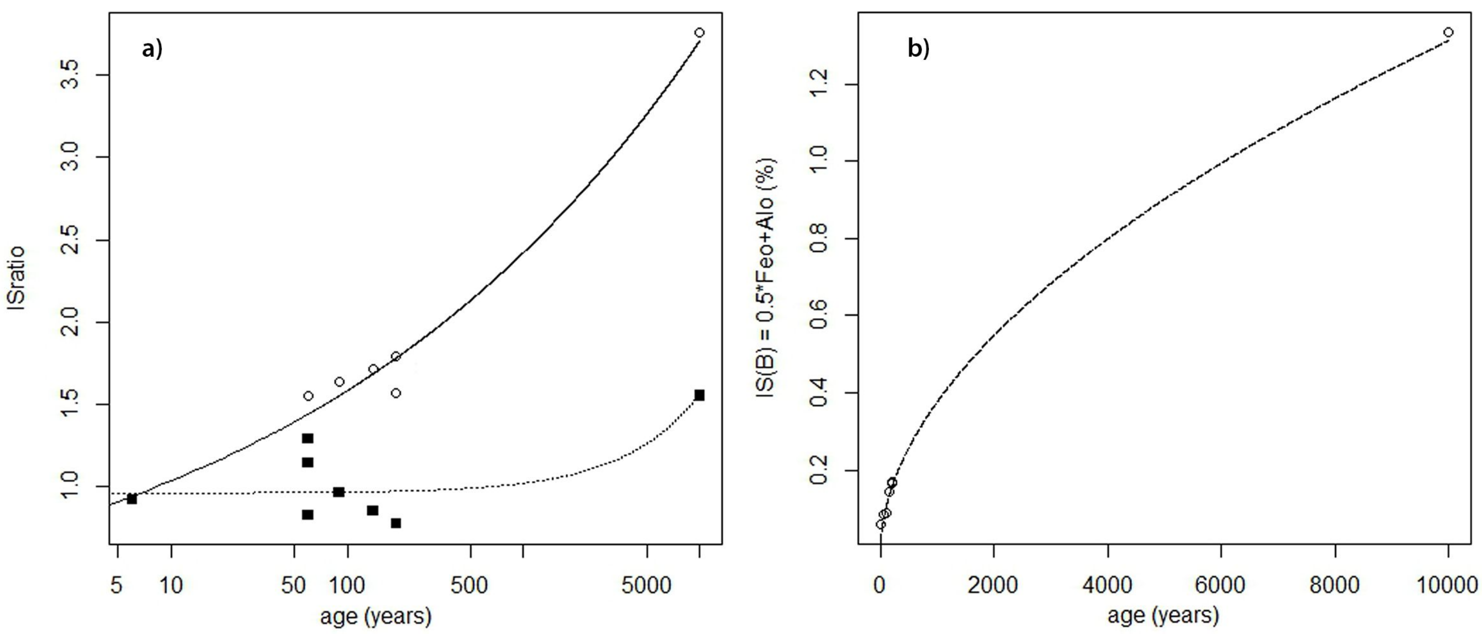
a)
b
a
C/N surface
10
15
20
25
Grassland
Forest
b)
b
a
pH, surface
4.0
4.5
5.0
5.5
6.0
6.5
Grassland
Forest
c)
b
a
O horizons cumulative thickness
0
5
10
15
Grassland
Forest
d)
b
a
log(Feo(B)/Feo(A/E))
0.0
0.5
1.0
1.5
2.0
2.5
Grassland
Forest
e)
b
a
log(Fed(subsurface) / Fed(surface))
-1
0
1
2
3
Grassland
Forest
f)
b
a
ISratio
1.0
1.5
2.0
2.5
3.0
3.5
Grassland
Forest



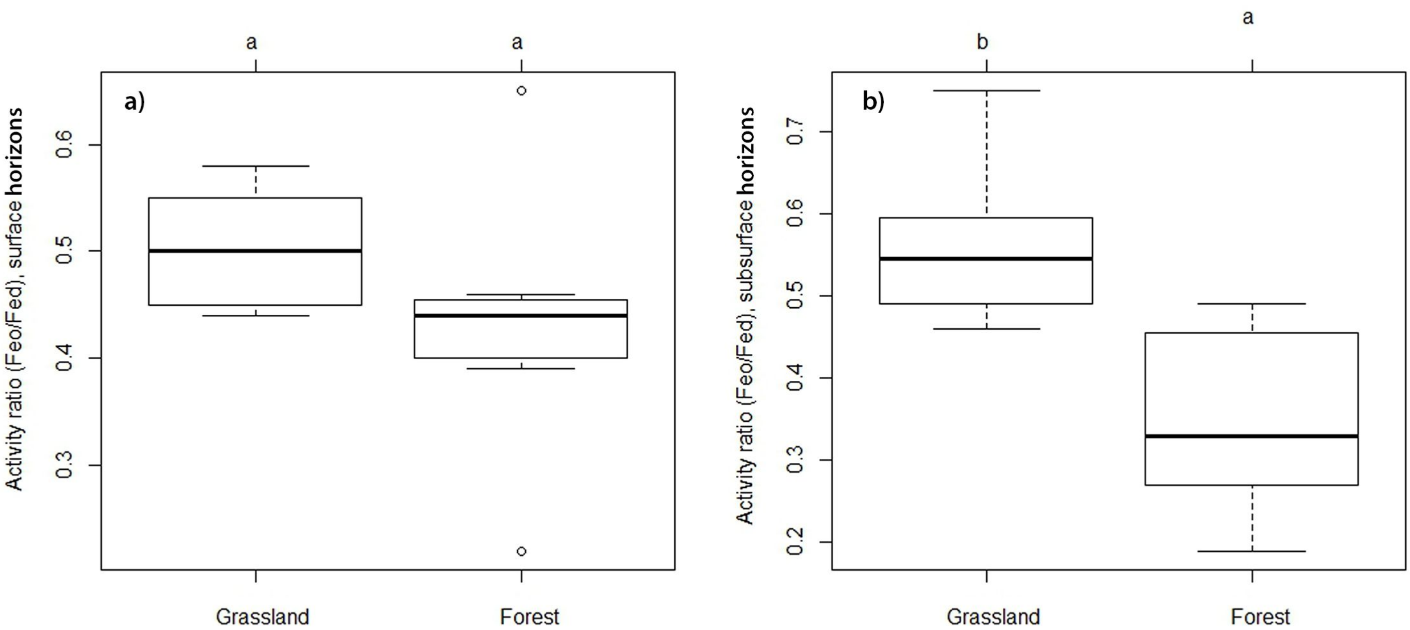
2004
c. 1950
1921
c. 1921
1821
1755
Site
Alpine (square)
Sub. grassland (circle)
Sub, typical (triangle)
Depths (cm)
Fe dithionite (mg/kg)



a)
b)
c)
d)
TOC (%, surface)
TOC (%), surface
TOC (%, subsurface)
TOC(subsurface) / TOC(surface) (%)
age
Grassland
Forest
a



a)
ISratio
3.5
3.0
2.5
2.0
1.5
1.0
5
10
50
100
500
5000
age (years)
b)
IS(B) = 0.5*Feo+Alo (%)
1.2
1.0
0.8
0.6
0.4
0.2
0
2000
4000
6000
8000
10000
age (years)



a
a
a)
Activity ratio (Feo/Fed), surface horizons
0.6
0.5
0.4
0.3
Grassland
Forest
b
a
b)
Activity ratio (Feo/Fed), subsurface horizons
0.7
0.6
0.5
0.4
0.3
0.2
Grassland
Forest