Hidden paleosols on a high-elevation Alpine plateau (NW Italy): evidence for Lateglacial Nunatak?

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Abstract

Alpine soils can provide valuable paleo-environmental information, representing a powerful tool for paleoclimate reconstruction. However, since Pleistocene glaciations and erosion-related processes erased most of the pre-existing landforms and soils, reconstructing soil and landscape development in high-mountain areas can be a difficult task. In particular, a relevant lack of information exists on the transition between the Last Glacial Maximum (LGM ~21,000 yr BP) and the Holocene (~11,700 yr BP), with this climatic shift that plays a crucial role for environmental thresholds identification. The present study aims at reconstructing the history and origin of hidden paleosols inside periglacial blockstreams and blockfields on a high-elevation Alpine plateau (Stolenberg Plateau) above 3000 m a.s.l., in the Northwestern Italian Alps. The results indicate that these soils recorded the main warming climatic phases occurred from the end of the LGM until the Late Holocene ~4,000 yr BP. Our reconstructions, together with the high carbon stocks of these paleosols, suggest that during warming phases the environmental conditions on the Plateau were suitable for plant life and pedogenesis, already since 22,000-21,000 yr BP. These paleosols reasonably evidence the existence of a Lateglacial Nunatak representing, to our knowledge, one of the first documented relict non-glacial

surfaces in the high-elevated European Alps. Thus, the Stolenberg Plateau provides important information about past climate and surface processes since the end of LGM, suggesting new perspectives on the long-term landscape evolution of the high European Alps.

Keywords: ^{14}C dating, $\delta^{13}C$, blockstream/blockfield, paleoclimate, Umbrisol, relict surface

Introduction

High-mountain areas can preserve traces of dramatic climatic variations, representing unique geosites and “storytellers" about the past landscape dynamics (Favilli et al., 2008). However, reconstructing soil and landscape evolution in such areas can be a difficult task because Pleistocene glaciations and related processes erased most of the pre-existing landforms and soils, leading to the formation of a complex mosaic of Quaternary sediments and soils of different ages (Sartori et al., 2001). Nevertheless, on scattered stable surfaces preserved during Pleistocene glaciations, ancient soils can be locally preserved for long periods (D’Amico et al., 2016). These soils, apparently in contrast with Holocene soil forming conditions, represent paleosols (when buried) or relict soils (Ruellan, 1971) that constitute excellent pedo-signatures of different specific past environmental conditions.

Relict surfaces are recognizable as flat summits and plateaus perched high above the valley floors at different elevations, in which erosion and deposition processes were very limited (D’Amico et al., 2016), because of lateral migration of glacial masses (Carraro and Giardino, 2004). Those surfaces that were not affected by the passage of glaciers experienced extreme cold conditions, which induced strong frost-action processes (e.g., frost-shattering, frost sorting, frost heave, etc.) (Karte, 1983), leading to the formation of periglacial features such as blockfields, blockstreams, and tors (Goodfellow, 2007; Ballantyne, 2010). Because of their high stability on certain poorly weatherable materials (D’Amico et al., 2019), these Pleistocene relict periglacial landforms are considered key indicators of ancient non-glacial surfaces (Goodfellow, 2007). Therefore, they have been used in paleoclimatic reconstructions, as their formation can be associated with specific past environmental conditions (e.g., Karte, 1983; Wilson, 2013; D'Amico et al., 2019). In particular, blockfields, which are usually associated with mountain summits and plateaus, have been used as paleo-indicators of nunataks (e.g., Ballantyne and Harris, 1994; Ballantyne, 1998, 2010) or non-erosive ice covers such as cold-based glaciers (e.g., Nesje et al., 1988; Kleman and Borgström, 1990; Hättestrand and Stroeven, 2002).

Nunataks (Dahl, 1987) are isolated hills or mountain peaks that projected above the ice shields and alpine-type icecap (Fairbridge, 1968). They have been proposed as possible biological refugia during glacial periods (Schönswetter et al., 2005; Goodfellow, 2007; Birks and Willis, 2008), serving as sources for the rapid reoccupation of the later deglaciated landscape (Fairbridge, 1968). While several studies have been focused on nunataks especially at high latitudes (e.g., Birks, 1994; McCarroll et al., 1995; Ballantyne et al., 1998; Kullman, 2008), there is a paucity of such works in the European Alps (Schönswetter et al., 2005; Carcaillet and Blarquez, 2017; Carcaillet et al., 2018), likely due to intrinsic difficulties in finding relict surfaces preserved from glaciations. Moreover, although many studies proposed paleoclimatic reconstructions in Alpine environments, they were mostly localized at elevation lower than 2200 m a.s.l. and in different climatic conditions (e.g., Kerschner and Ochs, 2008; Samartin et al., 2012a; Heiri et al., 2014). Furthermore, while the environmental changes from the Oldest Dryas to the Holocene are relatively well documented (e.g., Samartin et al., 2012b; Cossart et al., 2012, Heiri et al., 2014), a substantial gap of paleoclimate data during the transition between Last Glacial Maximum (LGM) and the Early Lateglacial still exists in the Alps, despite the well reported beginning of deglaciation, which occurred no later than 22,000-18,000 yr BP (Ivy-Ochs et al., 2006a; Ivy-Ochs, 2015, Monegato et al., 2017; Seguinot et al., 2018).

In 2017, on a high-elevation Alpine plateau (Stolenberg Plateau) covered by periglacial features (blockstreams and blockfields), at 3030 m a.s.l, thick and well-developed Umbrisols were detected. Despite the large stony cover and the extreme sparse vegetation, these soils showed carbon stocks comparable to alpine tundra or even forest soils (Pintaldi et al., 2021). Considering the high elevation, the specific morphology, the presence of large periglacial features and the well-developed soils within them, we hypothesize that the Plateau may represent an Alpine Nunatak on which ancient paleosols have been preserved. If these soils are truly paleosols, what their age could be? What is the origin of the soil organic carbon stored within them? If this soil organic carbon was not allochthonous (e.g. transported by wind or through atmospheric deposition), what kind of vegetation could grow at the time of soil formation? If the Plateau preserved very old paleosols, could it represent an Alpine Nunatak and what could have been the history of the past evolution of such high-elevated landscape?

Based on these considerations, this work aims at investigating the age and origin of soils discovered within blockstreams and blockfields (Pintaldi et al., 2021), through the application of plant remains investigation, plant fragments and soil ^{14}C dating, and soil $\delta^{13}C$ analysis. Moreover, based on the interpretation of the obtained results and literature data, the work aims at reconstructing the possible paleoenvironmental evolution of this high-elevated periglacial landscape since the end of the LGM.

2. Materials and Methods

2.1 Study Site

The Stolenberg Plateau (3030 m a.s.l.) is located at the foot of the southern slope of Monte Rosa (4634 m a.s.l.), along the border between Valle d'Aosta and Piemonte regions, NW Italian Alps (Long-Term Ecological Research-LTER site Istituto Mosso, 45°52'42.87''N, 7°52'0.64''E) (Fig. 1, Supplementary material 1). The Plateau has a south-east orientation and covers a surface of ca. 1.35 ha, with a slope angle between 0° and 13°. Meteorological parameters of the study area (air temperature and total liquid precipitation) were recorded by the Bocchetta delle Pisse Automatic Weather Station (AWS) (2401 m a.s.l., managed by ARPA Piemonte), located ca. 2.5 km east of the study area (on the same slope). The temperatures at the Plateau were obtained using the standard lapse rate of 6 °C km^{-1}. The mean cumulative annual snowfall was recorded by the Col d'Olen AWS (2901 m a.s.l., managed by the Italian Army, Comando Truppe Alpine - Servizio Meteomont), located ca. 500 m south-east of the Plateau. The LTER area has a total annual precipitation of ca. 1300±270 mm (1997-2019) at 2400 m a.s.l., with a winter minimum and a late spring-summer maximum. The Plateau has a mean annual air temperature of –2.4±0.7 °C (1988-2019) and a mean summer (June, July, August) air temperature of 4.4±1.3 °C; July is the warmest month, with a mean air temperature of 5.2±2.7 °C. The mean annual liquid precipitation is ca. 358±86 mm (1997-2019) while the mean cumulative annual snowfall is ca. 800±143 cm, with a snow cover lasting for at least 8 months (2008-2019).

The Plateau is covered by a thick layer of stones of variable size (from decimetric to metric), well organized in autochthonous blockfields, blockstreams/sorted stripes, gelifluction lobes, tilted stones, and weakly developed sorted circles (Pintaldi et al., 2021). No glacial striations or roches moutonnées have been detected on the few highly fractured rock outcrops. The parent material is composed of gneiss and mica-schists (Monte Rosa nappe, Pennidic basement), and metabasites (Zermatt-Saas unit) (Tognetto et al., 2021). The vegetation cover, which is almost absent or confined to small patches, reaching no more than 5% of the Plateau surface, is composed of alpine species such as *Silene acaulis, Carex curvula, Salix herbacea* in the vegetated patches, while *Festuca halleri, Poa alpina, Ranunculus glacialis, Leucanthemopsis alpina, Cerastium uniflorum, Oxyria digyna* and a few other scattered species grow also in the stone-covered area.

In 2017, during operational activities for constructing a new cableway station on the Plateau, three soil trenches were opened in the construction area close to the protected geosite (soil profiles P1, P2, and P3 in Fig. 1), revealing surprisingly well-developed soils under the stony cover. These soils were characterized by dark and thick organic C-rich A horizons (Fig. 2), and were classified as Skeletic Umbrisol (Arenic, Turbic), according to IUSS Working Group WRB (2015). Soil texture was generally loamy sand or sandy loam, pH (measured in H_2O) values were extremely to moderately acidic and carbonates were absent. Total Organic Carbon (TOC) content reached maximum values of ca. 20 g kg^{-1} in the A horizons of profiles P1 and P2 and over 10 g kg^{-1} in profile P3; the soil C stocks (up to ~ 5 kg m^{-2}) were comparable to vegetated or even forest soils, despite the extremely sparse vegetation cover (Pintaldi et al., 2021). Geophysical investigations indicated that these hidden soils were widespread on the Plateau. Moreover, no relevant permafrost bodies have been detected in the site, even though some permafrost patches cannot be excluded (Pintaldi et al., 2021). The detailed description of the soil profiles, as well as their physical and chemical properties, distribution, and thickness, are reported in Pintaldi et al. (2021). The ground surface thermal regime monitoring (2019-2020) showed no significantly negative soil temperatures under the blockstreams (Supplementary material 2, Fig. S1,2). However, during the snow-free season, soil temperatures under these periglacial features were colder than in nearby snowbed soils covered by vegetation (Supplementary material 2, Fig. S3, Tab. S1).

2.2. Plant remains analysis

The presence of few plant fragments mixed within the soil material was observed within soil samples collected from the umbric A horizons in the soil profiles. In order to isolate and identify the plant fragments within soil matrix, to reconstruct the possible past vegetation of the Plateau, a plant macroremains approach was adopted, starting from the assumption that plant material contained in paleosols may preserve the main features of soil "seed banks" (Ter Heerdt et al., 1996). Furthermore, the investigation was applied on two additional soil samples collected deep inside a soil-filled rock wedge (a vertical fracture in the substrate filled with vertically stratified soil materials, likely formed by freeze-thaw action), at 3 m depth (Fig. 3) along the southern border of the plateau (site “Wedge” in Fig. 1). We used the method for extraction of seeds and fruits from deep time sediments (e.g., Martinetto and Vassio, 2010), because recent experiences (Bertolotto et al., 2012) on a few soils showed that this was effective. Soil samples were processed in the laboratory with a very dilute solution of H_2O_2 (1%–3%), applied to disaggregate the biotic from the abiotic components and facilitate the floatation of the lighter and porous particles, usually fruits and seeds. Subsequently, the floating particles and the heavier materials that settled to the bottom were gently washed and sieved separately. After this material was dried and sorted by size, the fruits, seeds, and plant fragments were picked from the sieved residue. These were analyzed under a stereomicroscope and taxa were identified using atlases of recent fruits and seeds (Berggren, 1969, 1981; Bojňansky and Fargašova, 2007; Ercole et al., 2012); atlases of fossil fruits and seeds (Velichkevich and Zastawniak, 2006, 2009); and by comparison to the Modern Carpological Collection (MCC) at the Department of Earth Sciences at the University of Torino. Finally, these identifications were compiled into a database, and abundance data were generated based on fruit and seed counts.

2.3. Plant fragments and soil ^{14}C dating

Radiocarbon dating (^{14}C) was performed on six plant fragment samples, accurately selected after the plant macroremains investigation: five samples, consisting of recognizable but fossil-resembling plant fragments (dark coloured, mineral coatings), were obtained from soil samples collected in the umbric A horizons (usually in the 0-10 cm layer); one sample, consisting of unrecognizable plant fragments (strongly decomposed), derived from soil samples collected in the soil wedge at 3 m depth.

Furthermore, ^{14}C radiocarbon dating was performed on ten soil samples, nine of which selected from soil profiles: two from profile P1 (1,2), four from P2 (7,8,9,9bis) and three from P3 (2,3,5) (Fig. 1); one sample was collected in the only fully vegetated patch of the plateau (P4), at 10-20 cm depth (A horizon) (Fig. 1). The radiocarbon dating was performed at CEDAD, the Centre for Applied Physics, Dating and Diagnostics, Department of Mathematics and Physics “Ennio de Giorgi” - University of Salento, Lecce, Italy, using radiocarbon accelerator mass spectrometry (AMS) analysis

(Calcagnile et al., 2005) and the standard preparation methods (D'Elia et al., 2004). Radiocarbon dates of soil samples were calibrated in calendar age by using the software OxCal v. 4.4.4 Bronk Ramsey (2021) (Figs. S4-S8), based on atmospheric data (Reimer et al., 2020). Further methodological details can be found in Supplementary material 3. Radiocarbon dating of soils and sediments can be problematic due to the presence of pre-aged carbon (e.g., Lowe and Walker, 2000; Pessenda et al., 2001; Thorn et al., 2009) or fresh/allochthonous organic matter (Wang et al., 1996; Tonneijck et al., 2006). Therefore, terrestrial plant macrofossils have been considered the most reliable material for ^{14}C dating (Lowe and Walker, 2000; Hatté et al., 2001). However, the fossil record of the Alpine flora is generally scarce due to the lack of conditions suitable for the accumulation of macro-remains at the highest elevations (Lang, 1994). If materials such as charcoal, wood, or other plant macrofossils (Muhs et al., 2003) are lacking, dating of soils or sediments is generally accepted (Wang et al., 2014), especially in specific sites and under certain conditions (Lowe and Walker, 2000). Thus, the interpretation of ^{14}C dates must be adapted to the specific soil ecosystem under study (Tonneijck et al., 2006).

2.4. Soil $\delta^{13}C$ stable isotope signature

The soil $\delta^{13}C$ signatures are frequently used to reconstruct plant community history and the sources of soil organic carbon (Bai et al., 2012). To verify the typical $\delta^{13}C$ signature of present-day vegetated soils in the study area, which clearly reflects the existing vegetation (Meyer et al., 2014), soil samples were collected from the A horizons (~10 cm depth) of five vegetated permanent study sites at slightly lower elevation (Fig. S3) in the LTER site (2750-2900 m a.s.l.,): four from typical snow-bed community belonging to the *Salicion herbaceae* phytosociological alliance (site 1,2,6,8) and one from an alpine microthermal *Carex curvula*-dominated grassland, ascribable to the Caricion curvulae alliance (site 3). These soils were classified as Skeletic Regosols, Cambisols or Umbrisols (Magnani et al., 2017; Freppaz et al., 2019). The analysis was performed also on fourteen selected soil samples collected at the Plateau: two from profile P1 (1,2), five from profile P2 (7,8,9,9bis,13,15), five from profile P3 (1,2,3,4,5) and one from the vegetated patch P4 (10-20 cm depth, corresponding to the A horizon). The detailed scheme and description of sampling point within soil profiles under blockstreams/blockfields was described in Pintaldi et al. (2021). Samples were air-dried, sieved to 2 mm and checked using stereomicroscope to eventually remove macro-contaminants. Then samples were ground and sieved to 0.5 mm. The $\delta^{13}C$ signature of total organic carbon (due to the absence of carbonates) was directly determined using an Isoprime 100 continuous flow stable isotope mass spectrometer coupled to a Vario Isotope Select elemental analyzer (EA-IRMS; Elementar Analysensysteme GmbH, Hanau, Germany). The $\delta^{13}C$-values (‰) were calibrated relative to the international standard Vienna Pee Dee Belemnite (VPDB) by means of a three-point calibration using standard reference materials IAEA-600, IAEA-603 and IAEA-CH3. Measurement uncertainty was monitored by repeated measurements of internal laboratory standards and standard reference materials. Precision was determined to be ± 0.1 ‰ based on repeated measurements of calibration standards and internal laboratory standards. Accuracy was determined to be ± 0.1 ‰ on the basis of the difference between the observed and known δ values of check standards and their standard deviations. The total analytical uncertainty for $\delta^{13}C$ values was estimated to be ± 0.2 ‰. Significant differences (p-value < 0.05) in $\delta^{13}C$ values between present-day vegetated soils and soils under periglacial features were evaluated through one-way analysis of variance (ANOVA) combined with Tukey HSD test. Statistical analyses were performed using R software, v. 3.6.0 (R Core Team, 2019).

3. Results

3.1. Plant remains and ^{14}C dating

The quantity of plant fragments found in the soils was very scarce, representing a negligible fraction compared to the soil matrix. However, they were composed of remains with definite morphologies, including cm-sized leaves (consisting mainly of well-preserved and recognizable specimens of *Salix herbacea, Cerastium uniflorum* and *Poaceae* sp.) and mm-sized fruits and seeds, which have been mostly identified as belonging to taxa growing today in surrounding snowbed areas and in the small vegetated patch on the Plateau (Tab. 1, Fig. 1). Only in sample F, which was collected inside the wedge (Fig. 3), the degree of decomposition was much higher, so that the morphology of larger plant fragments was vague and not recognizable. Only a few tiny fruits and seeds still had diagnostic morphologies and allowed the identification of some plant taxa (Tab. 1)

Concerning radiocarbon dating, the plant fragment samples were all modern (after 1950 AD, Tab. 2), except the strongly decomposed sample F, which was dated back to 1,796-1,571 cal. yr BP (Tab 2, Fig. 4A).

3.2. Soil Organic Matter ^{14}C dating and $\delta^{13}C$ signature

Unlike plant fragments, the results from AMS on soil samples revealed a wide range of ages, covering several thousands of years and different well-distinct climatic periods, between ca. 22,000-21,000 yr and 4,400-4,100 cal. yr BP (Tab. 2, Fig. 4). All radiocarbon dates were rounded and here presented as calibrated radiocarbon ages (cal. yr BP = years before 1950 A.D.). In P1 the age of soil samples was related with depth, in fact the oldest samples P1-1, dated between 8,787 and 8,434 cal. yr BP (Tab. 2, Fig. 4B), was located close to the lower boundary of the umbric A horizon, while the youngest one (P1-2), dated between 5,744 and 5,583 yr cal. BP, was located close to the surface stones-soil interface (Fig. 2 A1-A2). In P2 the age of the samples was not related to depth: the oldest sample P2-9, dated between 17,536-17,014 and 18,228-17,870 (P2-9bis) cal. yr BP (Fig. 4C) (values obtained from two independent and blind datings performed in different moments), was located close to the surface stones-soil interface, while the stable organic C in C-rich cryoturbated patches (P2-7), located close to the lower boundary of the umbric A horizon, was much younger (6,500-6,306 cal. yr BP) (Fig. 2 B1-B2); the central and rather homogeneous part of the A horizon (P2-8) dates back to 8,534-8,302 cal. yr BP. In P3 the age of the samples was not related to depth as well: the oldest samples P3-2 and P3-3, dating back to 18,916-18,611 and 22,168-21,431 cal. yr BP respectively (Tab.2, Fig. 4D), were taken from homogeneous materials in the central part of the umbric horizon, while a younger radiocarbon age was obtained in the deeper part of the A horizon, close to the lower boundary (sample P3-5, Fig. 2 C1-C2), dating back to 13,337-13,110 cal. yr BP (Fig. 4E). The youngest soil sample, P4, taken from the A horizon in the currently fully vegetated patch, was dated between 4,405 and 4,090 cal. yr BP (Tab. 2, Fig. 4F).

The $\delta^{13}C$ signature of present-day vegetated soils provided by EA-IRMS ranged between -23.3 (Site 1) and -25.7 ‰ (Site 3), with a mean value of -24.2 ‰ (+/-1.1) (Tab. 3). The $\delta^{13}C$ of soil samples from the Plateau showed very similar values, ranging between -23.5 (P3-3) and -24.7 ‰ (P1-1), with a mean of -24.1 ‰ (+/- 0.4), while the soil sample collected from the vegetated patch (P4) had a slightly greater value of $\delta^{13}C$ of -22.7 ‰ (Tab. 3). No significant differences were detected between present-day vegetated soils and soils under periglacial features ($p<0.05$, Tukey HSD-test, Fig. 5).

4. Discussion

4.1 Plant fragments ^{14}C dating

The plant fragment samples were generally modern (Tab. 2), except sample F, which was dated between 1,796 and 1,571 yr BP, corresponding therefore to the small warm phase occurred during the Roman time (Mercalli, 2004). The strong difference in age between sample F and the other samples was already reflected in the different degree of decomposition, as the larger plant fragments in sample F were not recognizable. This sample, collected at ca. 3 m depth below the present-day surface inside a rock wedge, evidenced strongly active periglacial processes, sufficient to activate the rock wedge and thus its filling by surface soil material in the following centuries. This indicates that cryoturbation processes acted across millennia, strongly mixing plant fragments within the soil matrix, until they became unrecognizable. The presence of modern plant fragments within the soil matrix, although very scarce, can be explained mainly by the aeolian transport from the surrounding vegetated surfaces. The plant material could have been trapped by the stone layer and moved towards the soil surface through the large spaces between the rocks. Alternatively, a small input from the sporadically occurring vegetation growing within the stones may have contributed.

4.2 Soil ^{14}C dating

Differently from plant samples, datings of the soil samples collected from the Plateau covered an extended time interval, involving both Pleistocene and Holocene epochs. If some kind of very old-aged material (i.e. from LGM or even older) was deposited on the Plateau surface (i.e. derived from depositions on a cold-based glacier or by wind), it should have been deposited also on the nearby glacier surfaces. However, similar old-aged organic materials have never been detected in the well-studied Monte Rosa glaciers (Jenk et al., 2009; Thevenon et al., 2009). Currently, the oldest date obtained from an ice core, collected at ~80 m depth on the nearby Colle Gnifetti, was around 15,000 yr (Jenk et al., 2009). In the European Alps, on glacier surfaces, the most important source for the deposition of already aged organic materials is soil dust (Hoffman, 2016), a large part of which is originated from Saharan dust storm (e.g., Wagenbach and Geis, 1989; Wagenbach et al., 1996; Hoffman, 2016). The age of organic matter in these atmospheric depositions range between 1,000 and 5,000 yr (Eglinton et al., 2002; Jenk et al., 2006), with a mean of ca. 2,500 (Hoffman, 2016). Finally, the contribution of anthropogenic emissions was also considered as a possible cause for the aging of samples (Jenk et al., 2006), the so-called Suess effect (Suess, 1955). However, this effect is generally negligible on samples older than 2,000-5,000 yr BP (Graven et al., 2015; Köhler, 2016). Therefore, most of our samples, especially the oldest ones, were out of the range of influence of the Suess effect.

Based on the previous considerations, our soil samples cannot therefore be influenced by other than local organic carbon sources (i.e. vegetation grown during warming phases). Furthermore, the soil texture of the paleosols at the Stolenberg Plateau was loamy sand or sandy loam and no differences were found among profiles and between surface and deep samples (Pintaldi et al., 2021), thus rejecting also the hypothesis of a Loess deposition, which is otherwise mainly composed of silt-sized material dominated by quartz (Smalley et al., 2006), but could also include organic matter.

The surface position of the oldest samples and the general inversion of the typical age-depth relationship can be explained by the strong cryoturbation processes occurring on the Plateau, especially during the cold climatic phases. Indeed, cryoturbation processes mix and displace soil horizons (Bockheim and Tarnocai, 1998; Pintaldi et al., 2016), redistributing organic matter (e.g., van Vliet-Lanoë et al., 1998; Hormes et al., 2004; Bockheim, 2007). Furthermore, other works reported inverted soil age-depth relationship (e.g., Carcaillet, 2001; Favilli et al., 2008; Egli et al., 2009; Serra et al., 2020), as the young and contemporary carbon can be transported by soil turbation processes, such as cryoturbation and bioturbation, thus contributing to the rejuvenation of the subsoil (Scharpenseel and Becker-Heidmann, 1992; Rumpel et al., 2002; Favilli et al., 2008). Besides the well-developed periglacial features, cryoturbation processes were also evidenced by the internal soil morphology, which showed inclusions of surface A-horizon materials at depth, as well as strong convolutions and block displacement above wedges (Pintaldi et al., 2021).

Remarkably, ages similar to the ones of our oldest samples have never been detected in soils at such high-elevated ecosystems in the European Alps. For instance, Baroni and Orombelli (1996) found a Cambisol with buried A horizons at Tisa Pass (3200 m a.s.l.), but with a radiocarbon age of around 6,400-6300 yr BP, while Orombelli (1998) obtained an age up to ca. 9,000 yr BP (2500 m a.s.l.) for the Rutor Peat Bog. Glacier basal sediments at the Jamtalferner and Stubai glaciers (Austria) had ages around 17,000 and 22,000 yr BP, respectively (Hoffman, 2016). Furthermore, although at lower elevation (2100 m a.s.l.), Favilli et al. (2008, 2009) obtained comparable ages (17,000-18,000 yr BP) for an Entic Podzol, in the alpine belt in NE Italy. Other comparable radiocarbon ages (~21,000 yr BP) were reported by Carcaillet and Blarquez (2017) for a tree refugium at ~2200 m a.s.l., in the Western Alps.

Our soils could have experienced several different climatic conditions, retaining information about past climates, ranging from the end of the LGM (Ivy-Ochs, 2015) to the beginning of the Late Holocene (Walker et al., 2012), when a progressive cooling happened since ~4,000 yr BP (Deline and Orombelli, 2005; Orombelli et al., 2005). All the detected ages matched exactly and exclusively with the main warming phases/interstadials occurring from the LGM until the transition between the Holocene Climatic Optimum (HCO) and the Late Holocene (;), whereas no soil samples from cold phases/stadials were detected (further details in chapter 5).

4.3 Soil $\delta^{13}C$ signature

The overall $\delta^{13}C$ signature of present-day vegetated soils obtained by the stable isotope analysis, was comparable to those reported for soils and alpine vegetation in high-elevation ecosystems (Körner, 2003; Körner et al., 2016). The $\delta^{13}C$ signatures of soils under periglacial features corresponded very well with those of present-day vegetated soil (Tab. 3), thus indicating that the soil organic carbon probably originated from alpine plants with the same isotopic signature of present-day vegetation. As reported by Bai et al. (2012), when plant residues enter the soils, their $\delta^{13}C$ values may be modified slightly from their original values by isotope fractionation associated to preferential C mineralization. Furthermore, studies conducted by Colombo et al. (2020) on a nearby rock glacier at ~2700 m a.s.l., indicated $\delta^{13}C$ values of surrounding vegetated soils (-24.5 ‰) very similar to those of the Plateau, while the $\delta^{13}C$ signature of the active rock glacier soil, characterized by cold ground thermal regimes, coarse debris cover, and extremely reduced plant cover, increased considerably (ca. -18 ‰). Thus, the overall correspondence of the $\delta^{13}C$ signatures between present-day vegetated soils and paleosols under periglacial features suggested a common origin of the soil organic carbon from very similar alpine flora.

5. Historical and paleoenvironmental setting

In Fig. 6 we propose a conceptual model reporting a tentative paleoenvironmental reconstruction of the Stolenberg Plateau. Despite uncertainties, we believe that it may facilitate the interpretation of our data and also the generation (and testing) of the different hypotheses, as it includes and coordinates the different evidence we collected.

The LGM (Fig. 6A1-B1) ended around 22,000-19,000 yr BP (Ivy-Ochs et al., 2006a; Gianotti et al., 2015; Ivy-Ochs, 2015; Monegato et al. 2017), during which transection glaciers, flowing into valley systems, characterized the Western European Alps (Kelly et al., 2004). The mean air temperature was ~12 °C lower than present day in the European Alps (Peyron et al., 1998; Becker et al., 2016) and the mean July air temperature was likely around -4/5 °C on the Plateau (cf. Renssen et al., 2009; Samartin et al., 2012a,b; Heiri et al., 2015). Considering the lack of soil samples older than ca. 22,000 yr and the inferred cold conditions during LGM, together with the strongly weathered, autochthonous soil and stone materials, we can hypothesize the presence of a barren cryoturbated surface or of a small cold-based "ice cap" covering the Plateau (Fig. 6A1-B1).

Our oldest ^{14}C datings (P3-2, P3-3, P2-9, and P2-9bis) span from 22,000 to 17,000 yr BP, falling exactly during a period of massive downwasting of transection glaciers (Early Lateglacial Ice Decay-ELID) (e.g., Ravazzi, 2005; Ivy-Ochs et al., 2006a, 2008; Monegato et al., 2007; Reitner, 2007; Wirsig et al., 2016). Glacial shrinking also occurred at high elevations (Dielfolder and Hetzel, 2014), with some mountain peaks, around 2300-2600 m a.s.l., protruding out of the ice surface (Wirsig et al., 2016). The ELID occurred on both sides of the Alps due to significant rise in air temperatures (e.g., Huber et al., 2010; Schmidt et al., 2012; Samartin et al., 2012a,b). Assuming a pronounced climate continentality (Jost-Stauffer et al., 2001; Ivy-Ochs et al., 2009), summer air temperatures were likely similar to the ones inferred for the Bølling-Allerød interstadial (e.g., Schmidt et al., 1998, 2012; Huber et al., 2010; Samartin et al., 2012a,b). In addition, soil surface temperatures in alpine environments during summer are generally 2-4 °C above the air temperatures (Scherrer and Körner, 2010), indicating that life conditions of alpine organisms growing on the soil surface can be strongly decoupled from conditions in the free atmosphere, particularly on south oriented surfaces like the

Commented [F1]: South east?

Plateau (e.g., Scherrer and Körner, 2010). Our ancillary measurements confirmed that the mean 10 cm depth soil temperature during summer was ~3 °C warmer than air temperature on the Plateau (vegetated patch GST2), while at slightly lower elevation soil temperatures was 1-3 °C warmer (Supplementary material 2, Tab. S2). Therefore, the Stolenberg Plateau was likely ice-free since ~22,000-21,000 yr BP, i.e. the (micro)climatic conditions were probably suitable for pedogenesis and growth of some vegetation (Fig. 6A2,3-B2,3).

No radiocarbon ages were detected in our soils between ~17,000 yr BP and ~14,700 yr BP, which was a period (Gschnitz stadial or Oldest Dryas; Walker et al., 1999; Ivy-Ochs et al., 2006b, 2008) characterized by a decrease of both temperatures and precipitations, with values 8.5-10 °C and 25-50 % lower than the respective modern values (Ivy-Ochs et al., 2006a,b; Kerschner and Ivy-Ochs, 2008; Ivy-Ochs, 2015). These cold climatic conditions allowed a readvance of mountain glaciers (Kerschner et al., 2002; Ivy-Ochs et al., 2006a,b, 2008). Thus, it is possible to hypothesize that the environmental conditions on the Plateau might not have been suitable to sustain a sufficient plant life to produce enough detectable organic carbon, while they favored strong frost-action processes which led to the activation of periglacial features (Goodfellow, 2007; Ballantyne, 2010) (Fig. 6A4-B4).

The age of the P3-5 sample, dated between 13,337 and 13,110 cal. yr BP, matched perfectly with a warm period occurred between ~14,700 and 12,900 yr BP, corresponding to the Bølling-Allerød interstadial (Rasmussen et al., 2006; Ivy-Ochs et al., 2008; Dielfolder and Hetzel, 2014). A strong rise in the mean annual air temperatures was inferred (ca. 3 °C), with respect to the Oldest Dryas, causing the melting of valley glaciers (e.g., Ravazzi, 2005; Vescovi et al., 2007; Dielfolder and Hetzel, 2014). Other studies indicated even greater rises in temperature, ~3-4 °C (Larocque-Tobler et al., 2010) and ~5 °C (Renssen and Isarin, 2001). The mean July temperature at the Plateau, during this period, may have been higher than 3 °C (cf., Heiri and Millet, 2005; Samartin et al., 2012a,b; Dielfolder and Hetzel, 2014). Thus, the summer climate could have been suitable again for pedogenesis and plant life (Fig. 6A5-B5).

After the Bølling-Allerød, a general worsening of climate conditions occurred, leading to another cold phase, the Younger Dryas, also called Egesen Stadial (Ivy-Ochs et al., 2006b, 2008), which lasted until 11,700 yr BP. The summer temperatures were 3.5-4 °C lower, while precipitation was reduced by 10 to 30% compared to modern values (Kerschner et al., 2000, Kerschner and Ivy-Ochs, 2008). Again, on the Plateau, no soil organic matter radiocarbon ages were detected from this cold period, as the environmental conditions likely favored frost-action processes rather than pedogenesis, probably leading to a new expansion of periglacial features (Fig 6A6-B6).

The ages of four soil samples (P1-1, P1-2, P2-7, and P2-8) span from 8,800 to 5,600 yr BP, corresponding to the warm Holocene Climatic Optimum (HCO), occurred between 10,000 and 5,000 yr BP (Mercalli, 2004; Orombelli, 2011). In this period, a ~3-4 °C temperature increase was estimated with respect to the Younger Dryas (e.g., Tinner and Kaltenrieder, 2005; Ilyashuk et al., 2009; Larocque-Tobler et al., 2010; Samartin et al., 2012b). Glaciers were probably smaller than present day during the height of the HCO (e.g., Ivy-Ochs et al., 2009; Orombelli, 2011; Grämiger et al., 2018; Bohleber et al., 2020). Mean air temperature was up to 1-2 °C warmer with respect to the present-day values in the European Alps (e.g., Grove, 1988; Nesje and Dahl, 1993; Antonioli et al., 2000; Ivy-Ochs et al., 2009) and the inferred July temperature at the Plateau may have reached values around 6-7 °C (or even more) (cf. Ilyashuk et al., 2009; Samartin et al., 2012b), therefore above present-day values (cf., Birks and Willis, 2008; Ilyashuk et al., 2009; Samartin et al., 2012b). This likely led to conditions suitable for plant life (Fig. 6A7-B7).

The age of our youngest sample (P4), dated 4,405-4,090 yr BP, corresponded with a period of climate stability or slight cooling encompassed between 5,000 and 4,000 yr BP, after which a strong decrease in temperature was estimated (Heiri et al., 2015), which led to Alpine glacier expansion from 3,300 yr BP (Ivy-Ochs et al., 2009). After 3,300 yr BP, colder climatic conditions caused prolonged and frequent glacier advances, leading finally to the Little Ice Age (LIA, 1300-1850 A.D.) (Ivy-Ochs et al., 2009). During this last and prolonged cold phase, no soil radiocarbon ages were detected at the Plateau, apart from highly weathered plant fragments collected deep inside the rock wedge. The frost action likely prevailed, causing the final expansion of the periglacial features and the complete covering of the Plateau (Fig. 6A8-B8), while few plants could thrive without being able to leave measurable amounts of organic matter in the soil horizons.

6. Nunataks: yes or no?

The nunatak theory hypothesizes that unglaciated reliefs in glacial and periglacial areas acted as a refugium for isolated colonies of microorganisms, plants, and animals which survived the rigorous condition of the last glacial times for a few thousand years (Fairbridge, 1968; Dahl, 1987). These nunataks could have served as center for the recolonization of the later deglaciated landscape (Fairbridge, 1968). However, clear evidence for in situ survival of alpine floras on nunataks in the Alps during the last ice age are rather limited (e.g., Stehlik, 2002; Schönswetter et al., 2005; Carcaillet and Blarquez, 2017; Carcaillet et al., 2018) and subjected to a heated debate (e.g., Gugerli and Holderegger, 2001; Carcaillet and Blarquez, 2019; Finsinger et al., 2019). The existence and identification of such refugia during glacial or interglacial stages has been a topic of active research for decades (Hampe et al., 2013). The recolonization of the Alps would have started not only from peripheral refugia, but also from areas within the ice sheet (Schönswetter et al., 2005), where isolated nunataks could have been sources and targets as well of species immigration and establishment (Paus et al., 2006). Indeed, barren substrate or saprolite (Goodfellow, 2007), exposed just after glacier retreat (Fig. 6A2-B2), could become targets of autotrophic organisms (e.g., algae, mosses, lichens, higher plants), starting the process of primary succession (Bardgett et al., 2007). Thus, nunataks may have been indeed inhabited for several thousand years during the last glaciation (Gugerli and Holderegger, 2001), before the surrounding lowlands became deglaciated and invaded by organisms in the early Holocene. Remarkably, the Plateau location matched exactly with an area assumed to be a potential refugia for the survival of high-elevation plants on ice-free mountain tops within the strongly glaciated central parts of the Alps, particularly among the north of the Aosta Valley (NW-

Italy) and south Valais, and within the mountain ranges of Monte Rosa (Stehlik, 2002; Schönswetter et al., 2005; Kosiński et al., 2019).

Based on the results reported here and the presence of strong geomorphological evidences (i.e. periglacial features such as blockstreams/blockfields), as well as the overall specific morphology, aspect and position, the Stolenberg Plateau is thought to represent a Lateglacial Alpine Nunatak, on which specific pedoclimatic conditions could have been suitable for alpine plant life already since 22,000-21,000 yr BP. As sometimes observed at high elevation at present day (e.g. *Saxifraga oppositifolia* growing at 4500 m a.s.l. near the summit of Dom in Switzerland, Körner, 2011, or many other species up to ca. 4000 m a.s.l., Boucher et al., 2021), soil/substrate temperatures can be increased by solar surface warming in specific protected sites. This is particularly true where adiabatic winds, topography, local rock warming effect (Carturan et al., 2013), long-wave radiation from nearby rocky walls (i.e., the Mt. Stolenberg rock wall), and mass elevation effect (e.g. Monte Rosa Massif) (Samartin et al., 2012a), favor specific and stable microclimate features (e.g., Stewart and Lister, 2001), allowing the formation of the nunatak conditions.

6. Conclusion

In the severe periglacial environment of the Stolenberg Plateau, at 3030 m a.s.l., thick and well-developed Umbrisol were detected inside periglacial features (blockstreams/blockfields). As previously reported in Pintaldi et al (2021), these soils showed carbon stocks comparable to alpine tundra or even forest soils, despite the large stony cover and the scattered vegetation. Radiocarbon dating and soil $\delta^{13}C$ signatures indicated that these hidden soils were paleosols that recorded exclusively the main warming phases occurring since the end of LGM until the beginning of the Late Holocene cooling. This finding suggests that the environmental conditions on the Plateau were suitable for alpine plant life and pedogenesis, already since the end of LGM. Our results, coupled with the inferred paleoclimate reconstruction, indicate that the Stolenberg Plateau can be considered a direct evidence of a Lateglacial Alpine Nunatak, representing therefore a valuable natural and historical archive for unravelling the post-LGM history of the high-elevation landscape of the European Alps.

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Figures

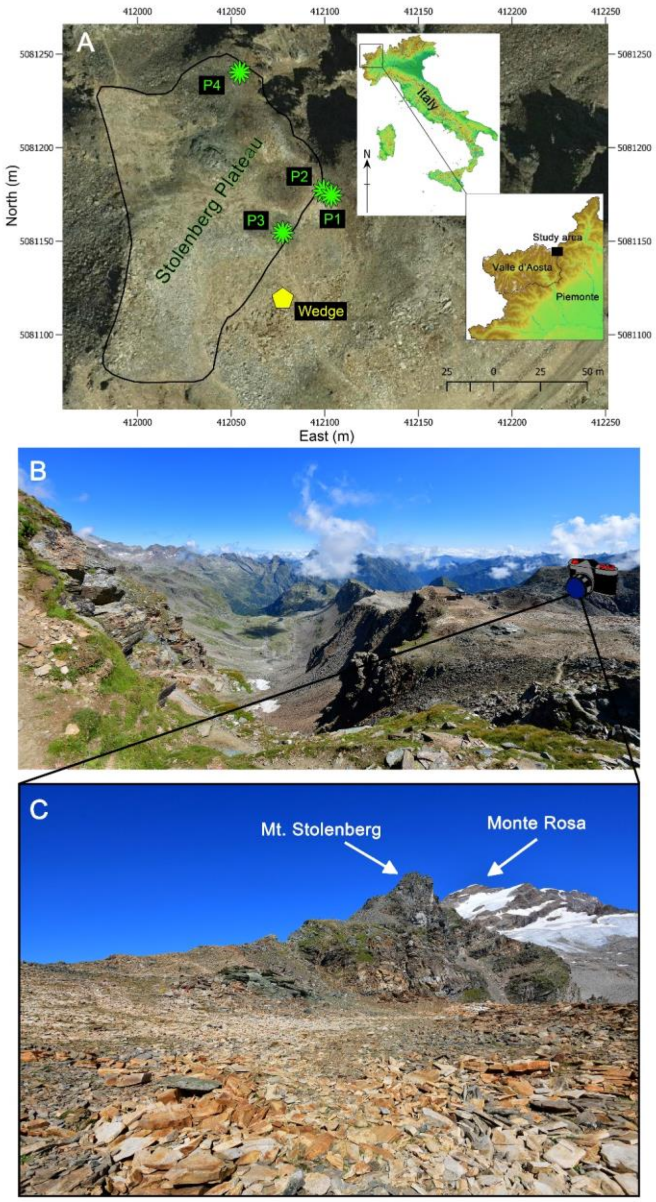

Figure 1. (a) Location of the study area in the NW Italian Alps (www.pcn.minambiente.it) and overview of the study area (orthoimage Piemonte Region, year 2010) (coordinate system WGS 84 / UTM zone 32N); green forms indicate the location of the three soil profiles (P1, P2, P3) and the vegetated patch (P4); yellow polygon indicates

the location of the soil-filled wedge. (b) View of the Plateau from the base of the Mt. Stolenberg (photo by M. D'Amico). (c) View of the Plateau (photo by M. D'Amico).

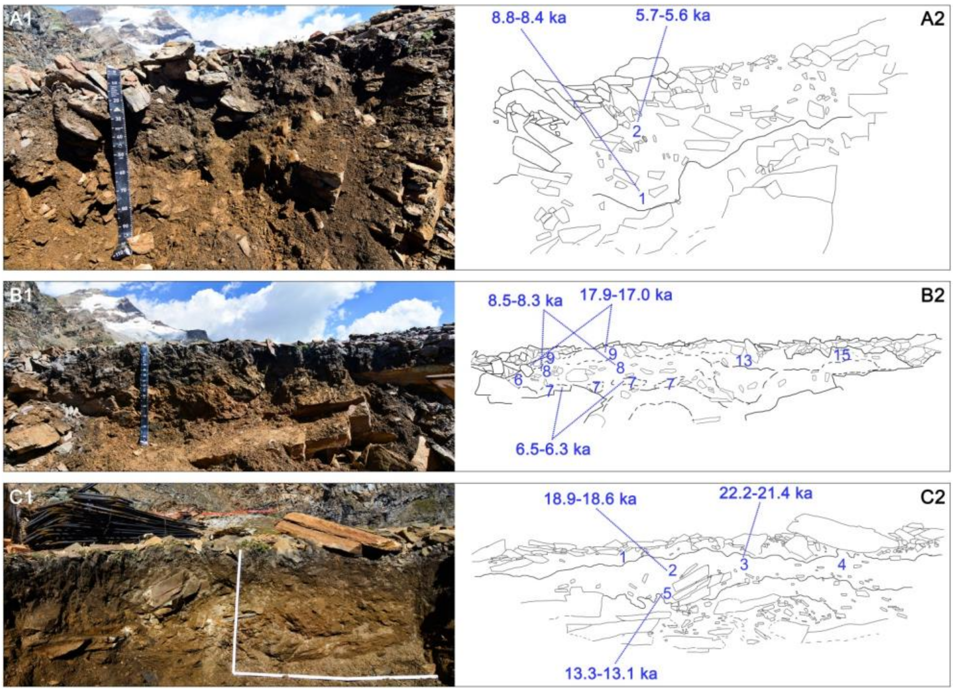

Figure 2. Soil profiles, with the corresponding scheme (on the side) reporting sampling points (number), the horizon limits (lines therein), and the age of soil samples (cal. ka BP). A1-A2: soil profile P1, samples P1-1 and P1-2 were analyzed for 14C (Tab. 2) and δ13C (Tab. 3); B1-B2: soil profile P2, , samples P2-7, P2-8, and P2-9 (and P2-9bis, not shown in the figure) were analyzed for 14C (Tab. 2); P2-7, P2-8, P2-9 (and P2-9bis), P2-13, and P2-15 were analyzed for δ13C (Tab. 3). C1-C2: soil profile P3, , samples P3-2, P3-3, and P3-5 were analyzed for 14C (Tab. 2); P3-1, P3-2, P3-3, P3-4, and P3-5 were analyzed for δ13C (Tab. 3).

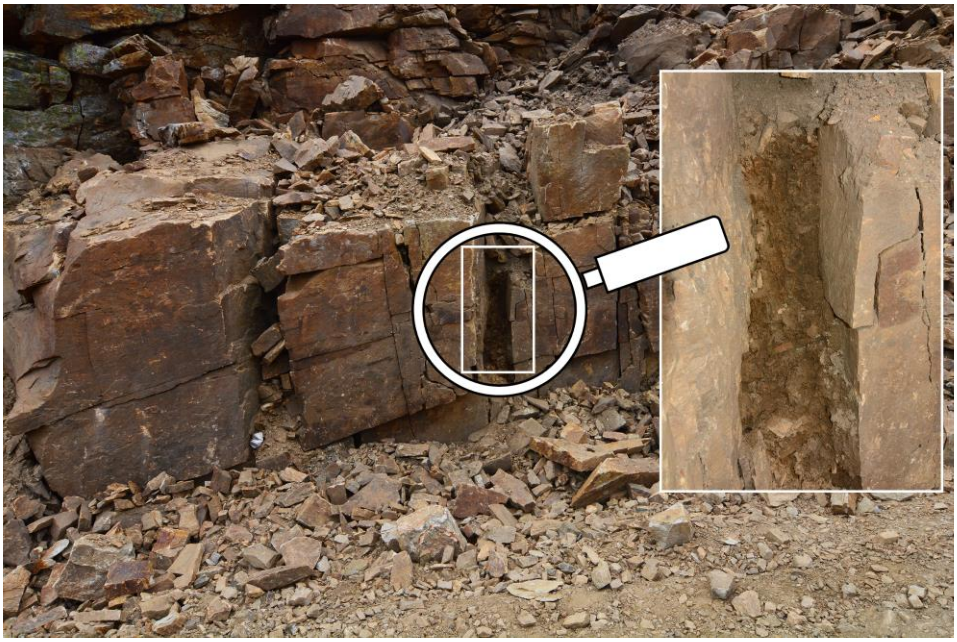

Figure 3. The soil-filled rock wedge along the southern border of the Plateau and detail of the sampling site.

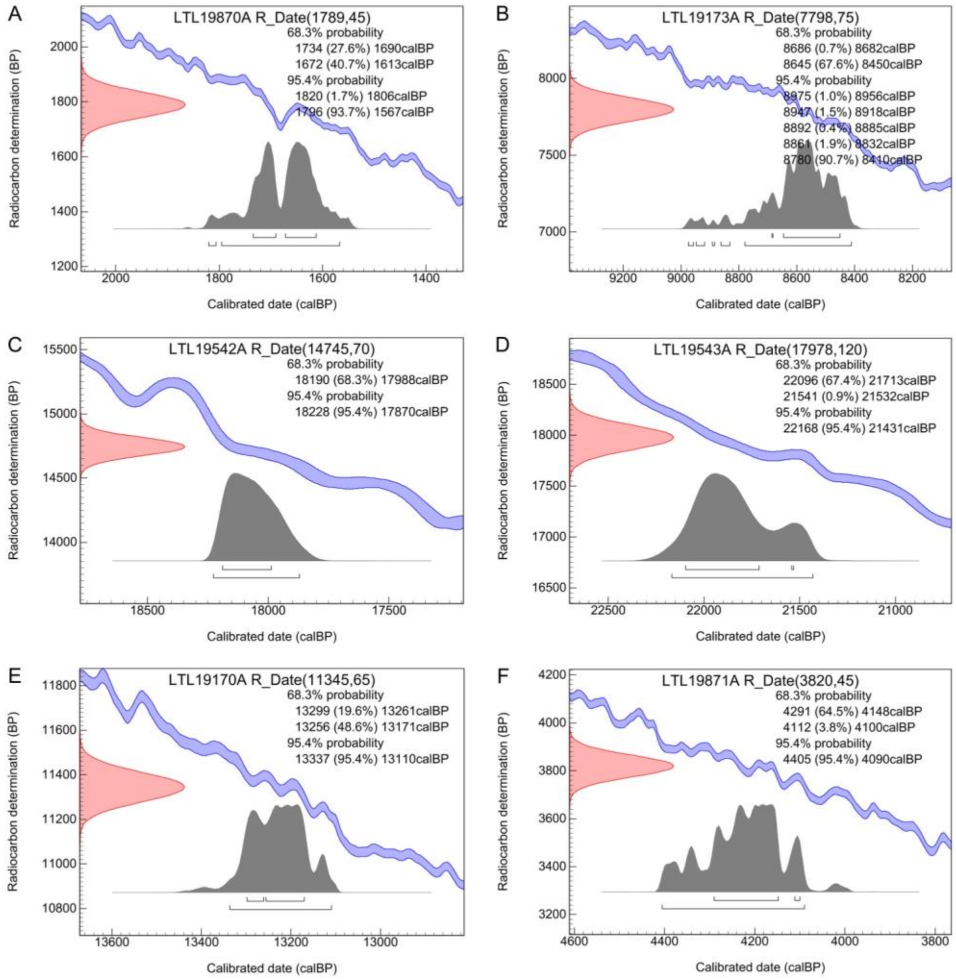

Figure 4. Radiocarbon date calibration of vegetation sample F (A) and soil samples P1-1 (B), P2-9bis (C); P3-3 (D), P3-5 (E), P4 (F). Radiocarbon date calibration of soil samples P1-2, P2-7,8,9 and P3-2 are reported in the Supplementary Material (Figs. S4-S8).

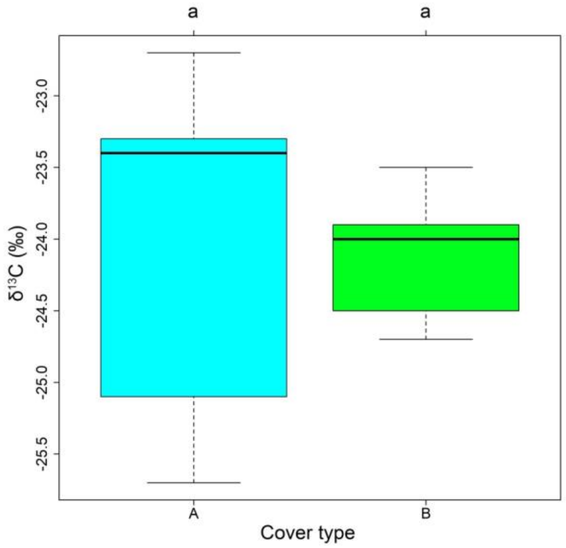

Figure 5. Boxplot (n=19) of $\delta^{13}C$ values of present-day vegetated soils (cover type A) and soils under blockstream/blockfield (cover type B). Letters (a) on the top indicate the absence of statistically significant differences among samples.

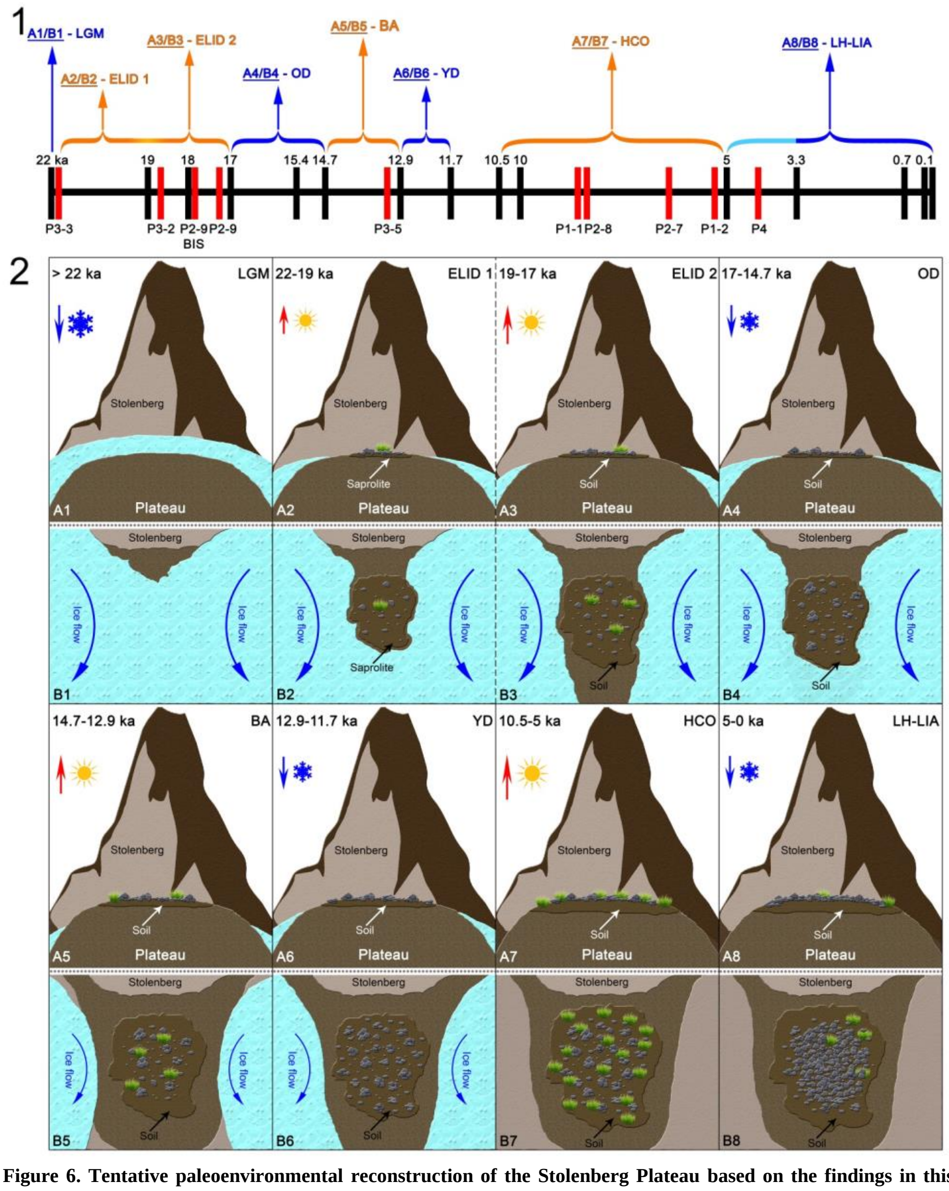

Figure 6. Tentative paleoenvironmental reconstruction of the Stolenberg Plateau based on the findings in this paper and on literature reported in the text. LGM: Last Glacial Maximum; ELID: Early Lateglacial Ice Decay; OD: Oldest Dryas; BA: Bølling-Allerød; YD: Younger Dryas; HCO: Holocene Climatic Optimum; LH: Late Holocene; LIA Little Ice Age. (1) Reference timeline from LGM to present day: blue colors indicate cooling phases, orange ones warming phases, the light blue segment (on the right) indicates a period of progressive cooling occurred between HCO and the LH phase; red segments indicate the age of soil samples reported in table 2. (2) Corresponding visual of the Plateau during the different phases: letters A (A1, A2, etc.) are the frontal views, B ones are the planimetric views. Details in the text.

Tables

Profile samples		
Family/Genus/Species	**N° (seeds/fruits/leaves)**	**Frequency (%)**
Asteraceae indet.	1	0.3
Brassicaceae indet.	1	0.3
Carex parviflora	1	0.3
Carex myosuroides	7	1.9
Cerastium sp.	16	4.4
Cerastium uniflorum	90	24.6
Cirsium sp.	1	0.3
Crepis sp.	1	0.3
Draba sp.	1	0.3
Gentiana gr. *verna*	1	0.3
Juncus sp.	1	0.3
Leucanthemopsis alpina	1	0.3
Minuartia sp.	23	6.3
Oxyria digyna	1	0.3
Poaceae sp.	75	20.5
Potentilla sp.	1	0.3
Salix cf. *herbacea*	103	28.1
Saxifraga oppositifolia	10	2.7
Sibbaldia procumbens	1	0.3
Silene acaulis	27	7.4
Taraxacum cf. *alpinum*	2	0.5
cf. *Vaccinium uliginosum*	1	0.3
TOT	366	100.0
Wedge sample (F)		
Family/Genus/Species	**N° (seeds/fruits/leaves)**	**Frequency (%)**
cf. *Artemisia*	1	5.9
Carex myosuroides	1	5.9
Cerastium sp.	1	5.9
Juncus sp.	1	5.9
Poaceae indet.	8	47.1
Primulaceae	1	5.9
Selaginella selaginoides	1	5.9
Silene acaulis	1	5.9
Taraxacum cf. *alpinum*	1	5.9
cf. *Vaccinium uliginosum*	1	5.9
TOT	17	100.0

Table 1. Results of the plant macrofossil investigation: identified plant taxa within soil samples collected from the Umbric horizons in the soil profiles and from the soil-filled rock wedge.

Lab. Code	Sample ID	TOC (g/kg)*	Type	Radiocarbon Age (yr BP)	Cal. Radiocarbon Age (cal. yr BP) (2σ range)	Phase
LTL19173A	P1-1	19.0	Soil	7798 ± 75	8787-8434*	HCO
LTL19174A	P1-2	10.8	Soil	4918 ± 45	5744-5583	HCO
LTL19175A	P2-7	20.5	Soil	5639 ± 45	6500-6306*	HCO
LTL19172A	P2-8	11.0	Soil	7608 ± 75	8534-8302*	HCO
LTL19176A	P2-9	11.3	Soil	14203 ± 100	17536-17014*	ELID
LTL19542A	P2-9bis	12.5	Soil	14745 ± 70	18228-17870	ELID
LTL19169A	P3-2	8.7	Soil	15463 ± 100	18916-18611	ELID
LTL19543A	P3-3	10.6	Soil	17978 ± 120	22168-21431	LGM/ELID
LTL19170A	P3-5	11.8	Soil	11345 ± 65	13337-13110	BA
LTL19871A	P4	13.8	Soil	3820 ± 45	4405-4090	LH
LTL19865A	A	-	Plant	modern	After 1950 AD	M
LTL19866A	B	-	Plant	modern	After 1950 AD	M
LTL19867A	C	-	Plant	modern	After 1950 AD	M
LTL19868A	D	-	Plant	modern	After 1950 AD	M
LTL19869A	E	-	Plant	modern	After 1950 AD	M
LTL19870A	F	-	Plant	1789 ± 45	1796-1571*	RWP

Table 2. Radiocarbon ^{14}C dating results of soil samples and plant fragments. HCO: Holocene Climatic Optimum; ELID: Early Lateglacial Ice Decay; BA: Bølling-Allerød; LH: Late Holocene; M: Modern; RWP: Roman Warm Period. *Total Organic Carbon (TOC) values derived from Pintaldi et al. (2021). *Weighted average age within the 95% confidence interval

Site	Elevation (m a.s.l.)	Cover type	$\delta^{13}C$ (‰)
S1	2840	Vegetation	-23.3
S2	2800	Vegetation	-25.1
S3	2770	Vegetation	-25.7
S6	2854	Vegetation	-23.4
S8	2749	Vegetation	-23.4
P4	3030	Vegetation	-22.7
P1-1	3030	Blockstream/Blockfield	-24.7
P1-2	3030	Blockstream/Blockfield	-23.9
P2-7	3030	Blockstream/Blockfield	-24.2
P2-8	3030	Blockstream/Blockfield	-23.9
P2-9	3030	Blockstream/Blockfield	-24.5
P2-9bis	3030	Blockstream/Blockfield	-24.5
P2-13	3030	Blockstream/Blockfield	-24.0
P2-15	3030	Blockstream/Blockfield	-24.3
P3-1	3030	Blockstream/Blockfield	-24.6
P3-2	3030	Blockstream/Blockfield	-23.8
P3-3	3030	Blockstream/Blockfield	-23.5
P3-4	3030	Blockstream/Blockfield	-24.0
P3-5	3030	Blockstream/Blockfield	-23.6

Table 3. IRMS $\delta^{13}C$ results of present-day vegetated soils in the study area and soils from the Plateau under blockstream/blockfield.