Effect of permafrost on the formation of soil organic carbon pools and their physical-chemical properties in the Eastern Swiss Alps

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17 **Abstract**

18 Current climatic conditions and the occurrence of discontinuous and sporadic permafrost in the Alps
19 result in a low turnover rate and therefore accumulation of organic matter (OM) in soils. Alpine soils
20 are thus highly sensitive to global warming that potentially promotes the mineralisation of soil organic
21 matter (SOM). This process might increase the release of CO₂ to the atmosphere. Our aim was to
22 investigate the potential effect of permafrost thawing by the analysis of the physical-chemical soil
23 properties of permafrost versus non-permafrost sites. Specifically, we i) quantified the SOM stocks at
24 such sites, ii) characterised SOM and its physical and chemical fractions and iii) estimated the age
25 range of the bulk soil and stable C-fraction (radiocarbon dating). In south-eastern Switzerland, two
26 areas above the timberline and one below the timberline (where isolated permafrost was verified) were
27 investigated in detail. At each site, the experimental set-up consisted in the comparison of nearby soils
28 that were either influenced or not by permafrost. The C-stocks (down to the C horizon or rock surface)
29 did not show a significant difference between permafrost and non-permafrost soils and were in the
30 same range of 10 – 15 kg/m² in alpine (grassland) and subalpine (forest) sites. Above the timberline,
31 the bulk SOM showed a distinct higher age at permafrost sites compared to non-permafrost sites. This
32 higher age was even more evident in the stable C-fraction (resistant to an H₂O₂ treatment), where ages
33 of up to 11 ky in permafrost soils were recorded. The highest age obtained in the stable C-fraction in
34 non-permafrost soils was around 4 ky. Consequently, climatic conditions and the occurrence of
35 discontinuous permafrost resulted in a very low turnover rate of SOM. At the subalpine site, the
36 difference between permafrost and non-permafrost sites was less. At both sites (alpine and subalpine),
37 DRIFT (Diffuse Reflection Infrared Fourier Transform) was used to determine the functional groups in
38 the bulk soil and in the stable C-fraction. In general, the stable C-fraction had a different composition
39 compared to the bulk SOM at non-permafrost sites; this was mostly not the case at the permafrost sites.
40 This confirms that different decomposition processes occur between permafrost and non-permafrost

41 sites. Furthermore, permafrost sites accumulated more the low-density physical fractions of SOM that
42 are potentially easily degradable.

43 The obtained results suggest that a warmer climate may not necessarily lead to an increased CO₂
44 release from SOM-degradation in permafrost soils compared to non-permafrost soils. High-alpine soils
45 and OM furthermore integrate a multi-faceted response to the past and ongoing surrounding
46 conditions. The melting of permafrost will most likely enhance vegetation growth, which to a certain
47 degree will probably compensate for carbon losses on the long-term.

48

49 Keywords: Alpine soils, climate, stocks of soil organic carbon, stability of soil organic carbon,
50 radiocarbon dating

51

52 **1. Introduction**

53 Permafrost is a widespread phenomenon in cold alpine and subarctic environments. Climate warming is
54 highly threatening the stability of permafrost soils in areas with mean annual air temperatures close to 0
55 °C. A thawing of permafrost in those sensitive ecosystems will not only enhance the risks regarding,
56 for example, slope stability; it will also increase the vulnerability of soil organic matter (SOM) to rapid
57 microbial decomposition that was previously stabilised by freezing temperatures (Kääb et al., 2007;
58 Schmidt et al., 2011).

59 Discontinuous permafrost distribution in the Alps is a function of altitude and exposure and can be
60 assumed to potentially occur at north-exposed sites above an altitude of 2400m asl, and on south-facing
61 sites above 3000m asl, respectively (Nötzli and Gruber, 2005). Due to the local microclimate, sporadic
62 permafrost can additionally exist below the timberline at very shaded sites (Kneisel et al., 2000). In the
63 Younger Dryas (late Pleistocene), permafrost in the south-eastern part of Switzerland had been present
64 in areas above 1950 m asl in north-exposed slopes, and above 2450 m asl on south-facing slopes (i.e.
65 500 m to 600 m below the present-day limits of discontinuous permafrost). The formation of Alpine
66 soils mostly started in the late Pleistocene or at the beginning of the Holocene and was characterised by
67 alternating warmer and cooler conditions. Some of these warmer inter-phases occurred during the
68 Older Atlanticum (5 to 8 ky BP) that is characterised as the ‘climatic optimum’ (Frauenfelder et al.,
69 2001; FOEN, 2005; Böhlert et al., 2011b). Permafrost distribution, soil genesis and SOM formation
70 was thus not always straightforward. Current soil properties integrate the effect of such climatic
71 variability.

72 A changing climate is generally assumed to affect soil organic carbon. What this might mean for soils
73 having a complex evolution is, however, largely unknown. The complex terrestrial carbon cycle is
74 typically controlled by three main factors: (i) climate (temperature and precipitation; Sanderman et al.,
75 2003) that is influencing vegetation establishment and microbial decomposition; (ii) the biochemical
76 composition of organic residues and SOM (Gleixner et al., 2001) which is highly dependent on plant

77 species and decomposition processes and (iii) the removal of organic carbon out of the carbon cycle —
78 or at least decelerated decomposition rates due to the stabilisation of organic matter, e.g. in organo-
79 mineral complexes (Krull et al., 2003).

80 The fractionation of SOM in its different decomposition and stability phases is essential because it
81 predetermines the vulnerability of SOM to environmental changes (Budge et al., 2011). Several
82 chemical and physical fractionation techniques exist to separate the labile C-fraction (only slightly
83 degraded OM) that mostly responds more rapidly to environmental changes, from the more stable,
84 recalcitrant mineral-associated C-fractions showing longer turnover times (Leifeld and Kögel-Knabner,
85 2005; Kögel-Knabner et al., 2008). The mechanisms involved in the stabilisation of the mineral-
86 associated soil carbon pools are still not entirely clear (von Lützow et al., 2006; Helfrich et al., 2007).

87 In permafrost environments mixing of soil layers due to cryoturbation can further influence the
88 formation of soil carbon pools: Freeze and thaw cycles enable the redistribution of organic carbon from
89 the surface to a greater soil depth or the other way around. The former process may retard its
90 decomposition and enables the soil to restart carbon accumulation in the topsoil layers (Ugolini et al.,
91 2006; Kaiser et al., 2007) whereas latter one will bring organic carbon to the surface horizons and
92 potentially may promote C-mineralisation.

93 Available data from alpine tundra soils show relatively large organic carbon concentrations in cold
94 alpine soils and a high abundance of less decomposed labile organic carbon with increasing altitude
95 (Jacot et al., 2000; Leifeld et al., 2009; Budge et al., 2011). It is generally assumed that permafrost soils
96 in arctic and subarctic lowland environments store a considerable amount of organic carbon, even
97 though possible mechanisms of carbon stabilisation by processes other than freezing are still unknown
98 (Hugelius and Kuhry, 2009; Schmidt et al., 2011). Most of the studies were done in high northern
99 latitudes with arctic and subarctic ecosystems (e.g. Bird et al., 2002; Zimov et al., 2006; Schuur et al.,
100 2008; Hugelius and Kuhry, 2009). Anoxic wetland conditions are typical for northern permafrost soils
101 and are responsible for the high accumulation of organic carbon. These specific processes cannot

102 directly be transferred to alpine permafrost soils that may differ in, for example, soil genesis, hydrology
103 and topography. Furthermore, only little data is available about the storage of organic carbon, its
104 chemical characteristics and turnover times in permafrost soils of the Alps. Consequently, the present
105 work focuses precisely on these issues by comparing permafrost soils with nearby non-permafrost soils
106 at the alpine and subalpine range in the south-eastern Swiss Alps.

107 The following research questions arose: (i) Are the SOM stocks at permafrost-sites really higher
108 compared to non-permafrost sites? (ii) Are the SOM characteristics and ages of C-fractions at such
109 sites different between permafrost and non-permafrost soils? (iii) Do the SOM stocks and chemical
110 characteristics differ between alpine permafrost and subalpine permafrost soils?

111 Our hypotheses for the study were: (i) permafrost soils contain a larger amount of labile organic matter
112 than non-permafrost soils, (ii) permafrost soils differ in their SOM quality and have a higher mean
113 residence time compared to non-permafrost soils, (iii) the subalpine permafrost sites experience a
114 warmer climate compared to alpine permafrost sites which should affect C-storage and C-chemistry.

115

116 **2. Materials and methods**

117 *2.1. Study area*

118 The study areas are located in the Upper Engadine (Switzerland) (Fig. 1). The Lateglacial and
119 Holocene history of the Upper Engadine area is documented in, e.g., Suter (1981) and Böhlert et al.
120 (2011b). The soils are skeleton-rich Leptosols (humic-skeletal) to Cambisols (WRB, 2006). The glacial
121 till deposited in these areas consists of granite/gneiss ('Julier Granite', Table 1). According to the Soil
122 Taxonomy (Soil Survey Staff, 2010), the soil moisture regime is udic (humid conditions, < 90 days/ per
123 year having a dry soil) at all sites and the soil temperature regime is cryic (mean annual temperature <
124 8 °C). Maximum precipitation occurs during the summer and autumn months. Two alpine and one
125 subalpine site were chosen (Fig. 1, Table 2):

126 1. Alpine areas (Bever and Albula):

These sites are situated at approximately 2700m asl. The climate is characterised by a mean annual temperature of -2.6 °C and a mean annual precipitation of >1000mm and according to the Köppen-Geiger classification defined as ET-climate (Kottek et al., 2006). At the Bever site, the occurrence of permafrost was measured using geoelectrical sounding. At both alpine sites the permafrost distribution was defined using an alpine permafrost map (Böckli et al., 2012). Permafrost soils are favoured at north-exposed slopes, whereas permafrost-free sites are found at south-facing sites. At the Bever site, permafrost features on the soil surface were further indicating cryoturbation caused by freeze and thaw cycles. The vegetation cover was predominantly grasses and herbaceous plants (*Caricetum curvulae*, *Geo montani-Nardetum*) and was usually denser at the permafrost-free sites, especially at the Bever site.

2. Subalpine area (Spinas):

This site is located below the timberline at 1800 m asl (at present the timberline is between 2200m and 2300m asl.), and represents one of the special places where isolated permafrost lenses could be confirmed by several geophysical techniques also at lower altitudes (Kneisel and Hauck, 2003; Kneisel and Schwindt, 2008). Mean annual temperature is 1°C and the mean annual precipitation is 1050 mm (Schwarb et al., 2000). According to the Köppen-Geiger classification, the climate is defined as Dfc (Kottek et al. 2006). Permafrost lenses and nearby sites without permafrost could both be found at north-exposed scree slopes due to differences in the microrelief and therefore site-specific climatology. There was no difference in the vegetation cover between the permafrost and non-permafrost soils. The forest is a *Larici-Pinetum cembrae*, with blueberry (*Vaccinium myrtillus*) and cranberry (*Vaccinium vitis-idaea*) in the herbaceous layer.

2.2. Experimental set-up

151 At each site (Bever, Spinass and Albula) 6 soil profiles (half of them having permafrost and half of them
152 without permafrost) were excavated. Approximately two to four kilograms of soil material (Hitz et al.,
153 2002) was collected per soil horizon. Soil samples down to the C horizon were taken. Soil bulk density
154 was determined by a soil core sampler. At the subalpine site Spinass, an additional set of 10 profiles
155 (down to a depth of 20 to 30 cm) was sampled to enlarge the database particularly with respect to soil
156 chemical properties. Half of these soils had permafrost and half were without permafrost.

157

158 *2.3. Soil Chemistry and Physics*

159 Total C and N contents of the soil were measured with a C/H/N analyser (Elementar Vario EL). Soil
160 pH (in 0.01 M CaCl₂) was determined on air-dried fine earth samples using a soil:solution ratio of
161 1:2.5. As the soils did not contain inorganic carbon (Julier granite is the parent material; cf. also Table
162 1), the total C contents equals the organic carbon content.

163 After a pre-treatment of the samples with H₂O₂ (3%), particle size distribution of the soils was
164 measured using a combined method consisting of sieving the coarser particles (2000 - 32 µm) and the
165 measurement of the finer particles (< 32 µm) by means of an X-ray sedimentometer (SediGraph 5100).

166

167 *2.4. Chemical fractionation of organic matter*

168 Our conceptual approach was based on the finding that partial oxidative degradation of OM leaves
169 behind intrinsically resistant as well as mineral-protected organic materials. This method is based on
170 the oxidation of OM by 10% H₂O₂ (Plante et al., 2004, modified; Eusterhues et al., 2005). 1 gram of
171 air-dried, untreated soil (< 2mm) was wetted for 10 min with distilled water in a 150-ml beaker.
172 Afterwards, 90 ml of 10% H₂O₂ were added. The procedure was run at a temperature of 50 °C and in a
173 closed systems throughout the treatment period. Peroxide treatments were performed for 168h (7 days).
174 At the end of the treatment the samples were washed three times with 40 ml deionised water and
175 freeze-dried, their weights recorded, analysed for total C and N content and ¹⁴C dated. Functional

176 groups were determined using DRIFT (Diffuse Reflection Infrared Fourier Transform). The amount of
177 organic C after the H₂O₂ treatment was related to the initial content using a mass-balance approach to
178 obtain the corresponding recoveries with

$$179 \quad \text{recovery of stable C (\%)} = (\text{gC}_{\text{after}} / \text{gC}_{\text{before}}) \times 100 \quad (1)$$

180 Recovery values were calculated also for nitrogen.

181 In the following, the OM that resisted to the H₂O₂ treatment is called the ‘stable C’. The soil fraction
182 oxidised by H₂O₂ is the ‘labile C’.

183

184 *2.5. Physical fractionation of organic matter*

185 The samples were air-dried. Large aggregates were gently broken by hand and sieved to < 2 mm. The
186 density fractionation procedure was adapted from Christensen (1992). Carbon fractions were separated
187 according to the following densities: <1.0 g cm⁻³ (without ultrasonic treatment) 1.0-1.6 g cm⁻³ (without
188 ultrasonic treatment), 1.0-1.6 g cm⁻³ (with ultrasonic treatment) 1.6-2.0 g cm⁻³ (with ultrasonic
189 treatment) and >2 g cm⁻³ (with ultrasonic treatment). The lightest fractions with a density <1 and 1-1.6
190 g cm⁻³ (without ultrasonic treatment, both fPOM) include the fresh or only slightly degraded organic
191 matter, but also inherent recalcitrant material such as charred residues. The fraction with a density 1-1.6
192 g cm⁻³ (with ultrasonic treatment, oPOM) represents carbon that is occluded within aggregates and
193 consequently physically protected for degradation. The fractions 1.6-2 g cm⁻³ and >2 g cm⁻³ (with
194 ultrasonic treatment, both mOM) are considered to be the more mineral associated C-fractions and
195 therefore chemically protected through interactions with mineral surfaces (von Lützow et al., 2006).

196 Forty grams of fine earth was added to 400 mL of deionised water, slightly stirred for 12 h, then
197 centrifuged for 10 min at 5000 g. The supernatant was sucked off and evaporated, so obtaining the <1 g
198 cm⁻³ fraction. The residual sample was dried at 40 °C and mixed to a sodium polytungstate solution
199 with a density of 1.6 g cm⁻³ to separate the 1.0-1.6 g cm⁻³ fraction from the >1.6 g cm⁻³ one. The
200 residual sample was again dried at 40 °C, mixed to a sodium polytungstate solution with a density of

201 1.6 g cm⁻³ but now an ultrasonic treatment of 50 J mL⁻¹ for 40 s was applied to the suspension, which
202 then was centrifuged for 30 min at 3000 g. The supernatant was sucked off and evaporated, so
203 obtaining the 1.0-1.6 g cm⁻³ fraction. The same procedure was finally applied to the residue using a
204 polytungstate solution with a density of 2.0 g cm⁻³ so as to separate the 1.6-2.0 g cm⁻³ from the >2 g
205 cm⁻³ one. All of the fractions were washed three times with deionised water, dried at 40 °C and
206 analysed for total C and N content.

207

208 2.6. DRIFT measurements

209 Relative peak intensities were used for DRIFT analysis (Bruker, Tensor 27). Spectra were recorded
210 from 4000 to 250 cm⁻¹ using a powder containing 9-12 mg of sample (3% of the total weight) and 291-
211 388 mg KBr (97% of the total weight). The samples were homogenised in a mill using a fine ball-mill
212 (Zr) for 45 s (frequency 25.0). Prior to measurement, the samples were dried in the oven at 70°C.

213 In order to quantify the relative changes in the FT-IR spectra, we divided the values of the relative
214 intensity (area) of each peak by the sum of the relative intensity and multiplied it by 100 using the
215 software OPUS 6. For the following peak range an individual peak search and calculation of relative
216 intensities (area) was done: (base1/base2)(cm⁻¹): 2980/2880, 1725/1710, 1660/1630, 1620/1600,
217 1535/1500, 1495/1470, 1470/1440, 1413/1333, 1190/1127, 1116/1050 and 1080/1030. The integration
218 method using a linear background between the found bases (individual samples) and absorbance values
219 was taken to calculate the relative concentration (OPUS 6). Major IR absorption bands and functional
220 groups assignments are given in Table 3. Aliphatic compounds were calculated using the IR range
221 1470-1440 cm⁻¹. The range between 2980-2828 cm⁻¹ was not considered because water may absorb in
222 this range and thus disturb the calculation procedure. DRIFT measurements were performed on the
223 H₂O₂ treated (stable C) and untreated soil material (bulk soil).

224

225 2.7. Carbon and nitrogen stocks

226 The total soil organic carbon (SOC) and stable carbon stocks were calculated down to the C horizon or
 227 rock surface according to the following equation:

$$228 \quad S_{stock} = \sum_i^{dz} C_i d_i \rho_i (1 - RM) \quad (2)$$

229 where S_{stock} denotes the abundance (kg/m^2) of the corresponding element and fraction, C the
 230 concentration of the element in the corresponding fraction (kg/t), d_i the thickness of layer i (m), ρ = soil
 231 density (t/m^3) and RM the mass proportion of rock fragments/soil skeleton.

232

233 *2.8. Radiocarbon dating of organic matter fractions*

234 Samples were cleaned using an acid-alkali-acid (AAA) treatment. The samples were then heated under
 235 vacuum in quartz tubes with CuO (oxygen source) to remove any absorbed CO_2 in the CuO. The tubes
 236 were then evacuated, sealed and heated in the oven at 900°C to obtain CO_2 . The CO_2 of the combusted
 237 sample was mixed with H_2 (1:2.5) and catalytically reduced over iron powder at 535°C to elemental
 238 carbon (graphite). After reduction, the mixture was pressed into a target and carbon ratios were
 239 measured by Accelerator Mass Spectrometry (AMS) using the tandem accelerator of the Laboratory of
 240 Ion Beam Physics at the Swiss Federal Institute of Technology Zurich (ETHZ). The calendar ages were
 241 obtained using the OxCal 4.1 calibration program (Bronk Ramsey, 2001; 2009) based on the IntCal 09
 242 calibration curve (Reimer et al., 2009). Calibrated ages are given in the 2σ range (minimum and
 243 maximum value for each).

244 Percent modern carbon (pMC) was calculated for obtaining the age of the labile organic matter
 245 fraction:

$$246 \quad pMC = 100 \times \left(e^{-\frac{(t+1950/1.03)}{8033}} \right) \quad (3)$$

247 The pMC of H_2O_2 -removable organic carbon (OC) was calculated from mixing models (Mikutta et al.,
 248 2006):

$$I - C_L = \frac{(I - C_U) \cdot UOC - (I - C_S) \cdot SOC}{LOC}$$

where $I - C_U$, $I - C_L$ and $I - C_S$ are, respectively, the values of untreated, H_2O_2 removable and H_2O_2 resistant (stable) OM. *UOC* (*untreated OC*), *LOC* (*labile OC*) and *SOC* (*stable OC*), correspond to the respective OC contents.

2.9. Statistics

The normal distribution of the data was verified with a Shapiro-Wilk test. A T-Test for data showing homogeneity of variance (F-Test) and Welch-Test for data not showing homogeneity of variance was performed in order to test the difference between two groups. For all tests a level of significance of 0.05 was considered. Statistical analyses were performed using the statistical software R (R development core team 2008, RVersion 0.95.263).

3. Results

3.1. General soil properties, SOM stocks and chemical fractions

All soils were acidic and the pH ranged from about 3.8 in the topsoil up to 4.6 in the deeper soil horizons (Table 4). The lower horizons and parent material consisted of loamy sand to sandy loam, whereas in the topsoil a higher silt and clay content could be measured giving rise to a sandy loam, a sandy clay loam or even a silt clay loam.

At the alpine and subalpine permafrost sites, average C-concentrations of 17 ± 3 g/kg (Bever; 0 – 30 cm depth), 54 ± 28 g/kg (Albula; 0 – 20cm depth) and 141 ± 88 g/kg (Spinass; 0 – 30cm depth), respectively, were measured in the bulk-topsoils. Non-permafrost soils at the Bever site showed a significant higher C-concentration (110 ± 30 g/kg) compared to the permafrost site, whereas the concentration at the Albula non-permafrost site (47 ± 20 g/kg) was similar to the permafrost soil. The subalpine site (Spinass)

273 indicated a (non-significant) trend towards a lower C-concentration in the permafrost soils (67 ± 29
274 g/kg) compared to the subalpine permafrost soils. With the exception of the Bever site, there was no
275 significant difference in the carbon concentration between permafrost and non-permafrost soils ($p <$
276 0.05).

277 After oxidation of the labile organic matter by H_2O_2 , the median stable C-recovery value for the alpine
278 area was near 13% for the permafrost soil (PF) and 9% for the non-permafrost soil (nPF). The
279 subalpine area had median recovery values of 9% (PF) and 10% (nPF), respectively (Fig. 2). In the
280 subsoils, where less labile organic matter is present, higher recoveries of the stable C-fraction were
281 measured (Table 4).

282 As found in previous investigations in the Alps (Favilli et al., 2008; Egli et al., 2010), the C/N ratio of
283 SOM dropped abruptly after the H_2O_2 -treatment (Fig. 2). The median C/N-value of the untreated soils
284 was near 9 (PF) and 10 (nPF) for the alpine area and 18 (PF) and 13 (nPF) for the subalpine area. After
285 the H_2O_2 -treatment, the C/N ratio (median) was only 2 (PF) and 1 (nPF) for the alpine site and 5 (PF)
286 and 4 (nPF) for the subalpine area. Compared to the untreated soils, the H_2O_2 -treatment thus leads to a
287 relative enrichment of nitrogen. In general, the C/N values obtained for the stable C-fraction (H_2O_2 -
288 resistant) are quite low. The C/N ratio of the labile C-fraction was consequently higher (Table 4). This
289 fraction is relatively enriched in carbon.

290 The calculated OM-stocks for the bulk soil, labile and stable C-fractions are given in Figure 3. At both
291 altitudes no distinct differences between permafrost and non-permafrost sites were detectable. For both
292 type of sites (alpine and subalpine), stocks of about 10 kg/m^2 for the permafrost and 15 kg/m^2 for the
293 non-permafrost soils were obtained.

294

295 3.2. Soil organic matter characteristics

296 The DRIFT spectra of the bulk SOM and its stable C-fractions (oxidised by H_2O_2) deliver further
297 information about its chemical characteristics.

298 In both areas (alpine and subalpine sites) the topsoil bulk organic matter consisted mostly of aliphatic
299 groups (data not shown), phenolic functional groups and, to a lesser extent, of polysaccharides and
300 secondary alcohols (Fig. 4). Alpine permafrost bulk soil was richer in secondary alcohols and phenolic
301 groups but showed a lower percentage of polysaccharides compared to the adjacent non-permafrost
302 bulk soil. A slightly higher abundance of secondary alcohols and polysaccharides but a lower
303 proportion of phenolic groups was measured in the subalpine permafrost bulk soil than in the alpine
304 permafrost bulk soil. Humification processes are related to the preferential oxidation of plant
305 polysaccharides and phenolic hydroxyl groups and are further characterised by an enrichment of
306 secondary alcohols and N-containing compounds in the stable C-fraction (Fig. 4). The relative
307 enrichment or depletion was constantly lower in the permafrost soils (alpine and subalpine sites),
308 pointing towards differences in the decomposition process when compared to the non-permafrost soils.

309

310 *3.3. Soil organic matter and density fractionation*

311 An overview of the C-concentrations (g/kg soil) of the different density fractions in the subalpine soils
312 is given in Figure 5. Different carbon concentrations were obtained in the individual density fractions
313 for permafrost and non-permafrost soils (subalpine site). The largest differences were measured in the
314 light C-fractions fPOM (<1 and $1 - 1.6 \text{ g/cm}^3$) and oPOM where permafrost soils contained about 59%
315 fPOM and 27% oPOM of the total carbon, respectively. At permafrost sites, the recovery of C was
316 constantly decreasing with increasing density (from $1 - 1.6$ to $>2 \text{ g/cm}^3$). Consequently, the lowest
317 percentages of carbon were detected in the mineral-associated fractions (mOM), showing values of
318 about 12% ($1.6 - 2.0 \text{ g/cm}^3$) and 2% ($>2.0 \text{ g/cm}^3$). In contrast, non-permafrost soils had only 42% of
319 fPOM and 16% of oPOM. In the non-permafrost soils, the mineral-associated C-fractions contributed
320 to a greater extent to the total carbon recovery with values of about 32% ($1.6 - 2.0 \text{ g/cm}^3$) and 10%
321 ($>2.0 \text{ g/cm}^3$), respectively. The C/N ratio decreased in both soils with increasing density, from mean
322 values of 28 and 26 (fPOM: $<1 \text{ g/cm}^3$) in permafrost and non-permafrost soils, respectively, down to

323 15 in permafrost and 12 in non-permafrost soils in the heaviest soil fraction ($>2 \text{ g/cm}^3$) (data not
324 shown).

325

326 3.4. Radiocarbon dating

327 Carbon-14 analyses were done for the bulk soil and stable C-fractions in the upper soil horizons of the
328 alpine and subalpine area. The age of these fractions are given in Table 5 for the alpine sites and in
329 Table 6 for the subalpine sites. Above the timberline, the bulk SOM in permafrost soils had a higher
330 age than the adjacent non-permafrost soils. Such differences were even more obvious in the stable C-
331 fraction in the permafrost soils where ages up to 3 ky (0 – 10 cm) and 11 ky (10 – 30 cm) were
332 measured, compared to a maximum age of 2.8 ky (0 – 10 cm), and approx. 4 ky (10 – 30cm) in non-
333 permafrost soils.

334 Below the timberline, the age of the bulk SOM (0 – 15cm) was similar between permafrost and non-
335 permafrost soils. In contrast, the stable C-fraction in permafrost soils (0 - 15 cm) was younger than in
336 the non-permafrost soils with a maximum age of up to 1.3 ky, whereas the stable C-fraction at non-
337 permafrost sites (0 - 15 cm) showed ages up to 3.6 ky. This pattern changed in the deeper soil horizons
338 (15 – 30 cm), where higher ages in permafrost bulk soils (up to 0.5 ky) compared to non-permafrost
339 bulk soils (modern age) were measured.

340

341 4. Discussion

342 4.1. SOC stocks and distribution

343 In the investigated alpine and subalpine soils, the stocks of bulk soil organic carbon (Fig. 3) was in the
344 lower range of previously reported results (Egli et al., 2008; Favilli et al., 2008; Hagedorn et al., 2010).
345 Hitz (2002) and Egli et al. (2008) found carbon stocks of up to more than 40 kg/m^2 in alpine soils with
346 frequent values between $15 - 25 \text{ kg/m}^2$. Hagedorn et al. (2010) measured carbon stocks of 15 kg/m^2 for
347 subalpine Swiss forest soils at similar altitudes. The variability in our data was, however, relatively

high. Carbon concentrations found in alpine and subalpine soils are consistent with values already reported for grassland and forest soils at those altitudes (Favilli et al., 2008; Egli et al., 2012). Compared to our investigated soils, arctic and subarctic lowland permafrost soils usually have higher organic carbon stocks. Hugelius and Kuhry (2009), for example, reported organic carbon stocks of 38 kg/m² (>100cm soil depth) in taiga and tundra permafrost soils of northeastern European Russia. In northern circumpolar permafrost soils, Tarnocai et al. (2009) even found carbon abundances of 32.2 – 69.6 kg/m² (0 – 100 cm soil depth). Compiled data for arctic, subarctic and boreal Cryosols revealed carbon stocks of 28.5 kg/m², 62.6 kg/m² and 77.6 kg/m², respectively in northern Canada (Tarnocai et al., 2000). Wetland conditions and cryoturbation, particularly in permafrost soils, are major soil forming processes in arctic and alpine regions and may contribute to a higher long-term storage of C in soils (Ugolini et al., 2006; Kaiser et al., 2007).

Surprisingly, the alpine grassland and subalpine forest sites showed similar organic carbon stocks (in permafrost and non-permafrost soils) in our investigations areas (Fig. 3). A higher primary productivity and consequently higher litter production and C inputs in the root zone were expected for the forested sites. The lack of difference might be explained by a higher litter input in the alpine zones and its temperature-limited conservation until present time and/or by a high productivity of grassland ecosystems. An additional possibility could be that the C input at high altitudes (with soils actually having permafrost condition) was greater during a certain period of time due to warmer climate and a different vegetation type.

The higher C storage capacity in soils having grassland is supported by a meta-analysis of Guo and Gifford (2002). They reported an equivalent or even higher potential of grassland ecosystems to store organic carbon than forest soils. Additionally, Tate et al. (2000) showed higher C-stocks (+13%) in the New Zealand tussock grassland than in a mountain beech forest (19.9 vs 16.7 kg/m²). Grasslands add organic matter in particular from below-ground (low shoot-to-root ratio) to the soil (Brown and Lugo, 1990). Hitz et al. (2001) even observed an increasing contribution of the dense fine root system to the

373 total carbon and nitrogen content and decreasing root turnover times in Alpine grasslands with
374 increasing elevation. Furthermore, Jackson et al. (2002) examined in a global analysis the relationship
375 of precipitation and to organic carbon storage in grassland and forest soils. They showed that grassland
376 productivity was strongly correlated to precipitation, with an increased total carbon content of around
377 43% compared to forest sites at precipitation rates of 1000 mm.

378 The climate experienced several variations during the late Pleistocene and Holocene and pedogenesis
379 has not been continuous and completely similar. A particularly warmer phase was recorded for the
380 Older and Younger Atlantic period (Maisch et al., 1999) when the treeline was 150-250 higher than the
381 present-day one (i.e. Burga, 1991; Favilli et al., 2010). Soil formation (chemical weathering) and the
382 storage mechanisms of organic C in the soil were consequently different to those of the present-day for
383 several millennia. This could mean that pedogenesis at the nowadays alpine sites occurred under a
384 different vegetation that had more similarities to the subalpine sites (more shrubs or even trees; Carnelli
385 et al., 2004). The uppermost tree limit reached 2500 ± 100 m asl in the Valaisan Alps during the
386 Holocene (Carnelli et al., 2004); i.e. it was 250 ± 100 m higher than today's potential tree line.
387 Consequently, the timberline would have reached 2400 ± 100 m asl. The absence of differences between
388 alpine and subalpine soils could be due to the palaeo-environmental effect.

389

390 4.2. SOC pools

391 In Alpine environments the highest C and N concentrations are usually found in the labile C- and N-
392 fraction, whereas the stable C- and N-fractions are only a minor part of the total C and N (Egli et al.,
393 2012). The relative proportion of the labile C-fraction was relatively high, ranging between 87 – 91%,
394 and similar at all investigated soils (PF and nPF) of the alpine and subalpine sites. The median recovery
395 value of the stable carbon fraction was consequently quite low (median: 9 – 13%, Fig. 2) but consistent
396 with values reported by other authors who measured a stable carbon fraction in the range of about 10 –
397 20% in Alpine soils having ages even up to 17 ky (Favilli et al., 2008; Egli et al., 2009; Egli et al.,

2012). It seems that the permafrost soils at the subalpine sites contained more weakly degraded organic matter than the alpine soils which gave rise to an increased C/N ratio (Table 4).

The stable C-fraction characterises an inaccessible, passive SOM pool and its stability is attributed to natural compound recalcitrance deriving from its biochemical characteristics, such as the presence of aliphatic macromolecules (lipids, cutans, algaenans, and suberans), charcoal and lignins and particularly to various processes of interaction with mineral surfaces (von Lützow et al., 2006). Compared to the bulk soil, we measured higher ^{14}C -ages in the stable C-fraction (Tables 5 and 6), a decrease of OH-deformation and C-O stretching of phenolic groups, an increase in the relative amount of proteic amide functional groups (N-H stretching of proteic amides) (Fig. 4) and a consistent decrease of the C/N ratio (Fig. 2). This is in agreement with recently reported results where it was hypothesised that protein-rich, most probably microbially processed, compounds have the capability to strongly adsorb onto mineral surfaces and are therefore crucial for the stability of organo-mineral particles (Kleber et al., 2007; Sollins et al., 2009). Also in the mineral-associated C-fractions (mOM: 1.6 – 2 and $> 2.0 \text{ g/cm}^3$) a lower C/N ratio was measured, pointing towards a lower C-loading with increasing particle density and an accumulation of proteinaceous material. Figure 4 furthermore gives the impression that in permafrost soils (alpine and subalpine), partially less degraded OM was able to attach directly to mineral surfaces (Fig. 4; see polysaccharides and phenolic groups).

The C-recovery after the H_2O_2 -treatment was consistent and low in all investigated soils (see chapter 3.1), but was not correlated to the percentage of carbon found in the heavier mineral-associated C-fractions (mOM: 1.6 – 2 and $> 2.0 \text{ g/cm}^3$) where a C-recovery of up to 42% was measured (see chapter 3.3). Von Lützow et al. (2006), for example, reported that density fractionation makes only a rough differentiation between an active and passive organic matter pool. While the light C-fraction is a good representation of the active pool, the heavy C-fraction contributes to the intermediate and passive pool. An explanation might be given by the model proposed by Kleber et al. (2007). They suggested a layered structure of organic matter attachment on mineral surfaces. A strong and direct interaction with

the mineral surface is favoured in situations where protein-rich material is adsorbed and forms a stable inner-layer zone ('contact zone'). Other layers, whose compounds are not directly attached to the mineral surface, surround this contact zone. The compounds in the outer layer are only loosely bound by, for example, cation bridging or hydrogen bonding, which leads to a shorter residence times for the individual molecular fragments. The thickness of this outer region depends more on the material input than on the availability of binding sites (Kleber et al., 2007). Accordingly, we hypothesise that the chemical treatment (H₂O₂) was better suitable to isolate the stable inner contact zone whereas the physical fractionation method (density fractionation) cannot separate layers of different bonding-strength.

432

433 4.3. SOM quality

In alpine soils, the relatively small-scale heterogeneity of the vegetation (see chapter 2.1) reflected the impact of the microtopographic position and microclimate (influence of snow cover, soil moisture etc; Haugland and Beatty, 2005). The bulk SOM at alpine permafrost and alpine non-permafrost sites differed slightly in its chemical composition (Fig. 4). In alpine permafrost bulk soils a higher abundance of secondary alcohols, phenolic groups and proteic amides, but a lower percentage of polysaccharides were measured compared to alpine non-permafrost bulk soils. Additionally, the stable C-fraction in alpine permafrost soils was less enriched in those compounds. These findings indicate differences either in plant species and/or decomposition processes between alpine permafrost and alpine non-permafrost soils. It is well known that the amount of residue input depends on the vegetation productivity, whereas SOC quantity and quality is largely determined by plant species (Helfrich et al., 2006) and the extent of decomposition that is related to the microbial community (Ohtonen et al., 1999; Kögel-Knabner et al., 2002). We hypothesise therefore that the differences in the chemical composition of SOM between alpine permafrost and alpine non-permafrost soils can be related to a lower C input, a reduced microbial activity and associated decomposition process in alpine

448 permafrost soils. Differences in the decomposition conditions are further supported by the radiocarbon
449 dating of the alpine bulk soils and its stable C-fractions. ^{14}C -ages of bulk soils and related labile and
450 stable C-fractions (Table 5) were distinctly higher in alpine permafrost soils than in alpine non-
451 permafrost soils. This indicates a higher mean residence time and therefore a lower organic matter
452 turnover. Also Favilli et al. (2009) reported high ages of the resilient C-fraction, often up to 11-14 ky
453 (an in one case up to 17.5 ky; 4-8 cm depth) for topsoils in the subalpine region (close to the
454 timberline) and up to 11 ky for topsoils in alpine soils. Radiocarbon dating of the stable C-fraction can
455 also be used as a surface age indicator (i.e. Favilli et al., 2008). Assuming that both soils (alpine
456 permafrost and alpine non-permafrost) started to develop at the same time, the higher ages obtained in
457 the stable fraction of alpine permafrost soils are consequently reflecting a better stabilisation due to
458 permafrost conditions.

459 Contrary to the alpine sites, no difference in the vegetation type and cover was visible between
460 subalpine permafrost and subalpine non-permafrost soils (see chapter 2.1). In the Ah horizon of these
461 permafrost soils, higher carbon concentrations were found than in the respective horizon of the non-
462 permafrost soils. In addition, subalpine permafrost bulk soils showed a higher abundance of secondary
463 alcohols and polysaccharides and a decrease of the relative proportion in phenolic groups compared to
464 subalpine non-permafrost bulk soils (Fig. 4). Due to the higher amount of carbon recovered in the
465 lightest soil fractions (fPOM, oPOM) in subalpine permafrost soils, we suggest that subalpine
466 permafrost soils hold a higher amount of labile C while decomposition is more favoured in subalpine
467 non-permafrost soils. The assumption of a lower biodegradation in the subalpine permafrost bulk soil is
468 further supported by its high C/N ratio that indicates a higher amount of fresh or only weakly
469 decomposed organic matter (Sollins et al., 2006) compared to subalpine non-permafrost soils.

470 In the upper 15 cm of the subalpine soils, the ^{14}C -ages obtained in the subalpine bulk soils were similar
471 between permafrost and non-permafrost soils, whereas the stable C-fractions in permafrost soils were
472 younger compared to the non-permafrost site (Table 6). This was rather surprising and did not support

our hypothesis that older SOM should also be found in the permafrost soils at subalpine levels. The reasons for this are still not clear. It appears that a better stabilisation of SOM with mineral surfaces or inclusion into aggregates was achieved in the subalpine non-permafrost soils. These processes are probably disrupted in the subalpine permafrost soils. From 15 to 30 cm soil depth, the subalpine permafrost bulk SOM, however, showed higher ^{14}C -ages than the subalpine non-permafrost soil. This, in turn, would support our hypothesis that decomposition processes are slower in permafrost soils.

479

4.4. *Climate warming and potential changes*

The response of decomposition and net-primary production (NPP) to climate warming is not balanced. Decomposition rates are changing faster as a function of temperature than does net-primary production (NPP) (Schimel et al., 1994). This imbalance in the temperature sensitivity between NPP and SOM decomposition predicts that SOM storage should decrease in response to climatic warming (Lützow and Kögel-Knabner, 2009), at least in the short term.

This is only in part supported by our results: we obtained higher ages for the bulk C and labile C-fraction in permafrost soils (Tables 5 and 6; Fig. 6). Higher ages can be due to a lower input of C to the soil and/or lower decomposition rates (that finally also depend on the communities of macro- and micro-decomposer; Eskelinen et al., 2009; Seeber et al., 2009; Leifeld et al., 2012). In the stable C-fraction, the ages follow, with some exceptions, better the 1:1 line (Fig. 6). The climate seems to be an important driver for the bulk C and labile C-fraction (and less for the stable C-fraction).

In addition, differences in SOM quality (Fig. 4) and the higher contribution of potentially easily degradable fPOM and oPOM fractions (lighter soil fractions, Fig. 5) to total SOM in permafrost show that a climate warming could give rise to an increased microbial decomposition. However, the carbon stocks in the bulk and labile fractions of the investigated permafrost soils were not significantly higher than in non-permafrost soils. It is therefore questionable whether a thawing of permafrost really would lead to an accelerated and increased carbon loss in these soils. For example, several scenarios of global

498 change are predicting ascending vegetation zones, with the subalpine coniferous forest and Ericaceous
499 shrubs becoming able to colonise meadows at higher altitudes (Ozenda and Borel, 1991). This shift in
500 vegetation is expected to increase primary production, but how such a shifting will influence the carbon
501 storage in the rooting zone is much more difficult to predict. Changes in plant species and a potential
502 increase in vegetation growth will enhance the aboveground carbon storage capacity that might offset
503 initial carbon losses.

504

505 **5. Conclusions**

506 We compared soils that have developed at sites either with or without permafrost conditions at a
507 subalpine and alpine level. We obtained the following main findings:

- 508 - Astonishingly, the organic carbon stocks did not significantly differ between permafrost and
509 non-permafrost sites. In addition the stocks were similar at the subalpine (forest) and alpine
510 (grassland) zone.
- 511 - Major differences between permafrost and non-permafrost influenced soils were however
512 detected in the chemical and physical characteristics of soil organic matter (SOM). The
513 chemical composition of SOM indicated differences in decomposition and stabilisation
514 mechanisms.
- 515 - The permafrost soils had a higher contribution of the free-particulate organic matter (fPOM)
516 and occluded organic matter (oPOM) to the total SOM. These fractions are potentially easily
517 degradable.
- 518 - In some permafrost soils a very high age of the stable SOM (resistant to H₂O₂ oxidation) could
519 be measured. At the alpine sites, higher ¹⁴C ages of the bulk SOM and stable C-fraction reflect
520 lower SOM decomposition rates and a very long-term fixation of the stable C-fraction. At the
521 subalpine sites, rather opposite processes seem to occur that cannot as yet be explained.

522 The findings are not all equivocal and demonstrate that Alpine soils had a complex soil genesis that
523 now also affects their future behaviour (under the light of climate change). Carbon accumulation and C
524 fractionation occurred over a longer period (c. 5 – 8 ky BP) in a warmer climate that still imprints soil
525 characteristics to a certain degree. It is not really evident that the permafrost soils would release more
526 CO₂ with a warmer climate. Furthermore, a potential increased release might be offset due to the
527 shifting dominance from grassland to shrub/woodland.

528

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532

533 **References**

- 534 Bird, M.I., Santruckova, H., Arneth, A., Grigoriev, S., Gleixner, G., Kalaschnikov, Y.N., Lloyd, J.,
535 Schulze, E.-D., 2002. Soil carbon inventories and carbon-13 on a latitude transect in Siberia. *Tellus*
536 54B, 631–641.
- 537 Boeckli, L., Brenning, A., Gruber, S., Noetzli, J. 2012: Permafrost distribution in the European Alps:
538 calculation and evaluation of an index map and summary statistics. *The Cryosphere*, 6, 807–820.
- 539 Böhlert, R., Mirabella, A., Plötze, M., Egli, M., 2011a. Landscape evolution in Val Mulix, eastern
540 Swiss Alps – soil chemical and minerlogical analyses a sage proxies. *Catena* 87, 313-325.
- 541 Böhlert, R., Egli, M., Maisch, M. Brandová, D., Ivy-Ochs, S., Kubik, P.W., Haeberli, W., 2011b.
542 Application of a combination of dating techniques to reconstruct the Lateglacial and early Holocene
543 landscape history of the Albula region (eastern Switzerland). *Geomorphology* 127, 1-13.
- 544 Bronk Ramsey, C., 2001. Development of the radiocarbon calibration program OxCal. *Radiocarbon* 43,
545 355-363.
- 546 Bronk Ramsey, C., 2009. Bayesian analysis of radiocarbon dates. *Radiocarbon* 51, 337-360.

547 Budge, K., Leifeld, J., Hiltbrunner, E., Fuhrer, J., 2011. Alpine grassland soils contain large
 548 proportions of labile carbon but indicate long turnover times. *Biogeoscience* 8, 1911-1923.

549 Brown, S., Lugo, A.E., 1990. Effects of forest clearing and succession on the carbon and nitrogen
 550 content of soils in Puerto Rico and US Virgin Islands. *Plant and Soil* 124, 53-64.

551 Burga, C.A., 1991. Vegetation history and paleoclimatology of the Middle Holocene: pollen analysis of
 552 alpine peat bog sediments, covered formerly by the Rutor Glacier, 2.510 m (Aosta Valley, Italy).
 553 *Global Ecology and Biogeography Letters* 1, 143-150.

554 Carnelli, A.L., Theurillat, J-P., Thinon, M., Vadi, G., Talon, B., 2004. Past uppermost tree limit in the
 555 Central European Alps (Switzerland) based on soil and soil charcoal. *The Holocene* 14, 393–405.

556 Christensen, B.T., 1992. Physical fractionation of soil and organic matter in primary particle size and
 557 density separates. *Advances in Soil Science*, vol. 20. CRC Lewis Publishers, Boca Raton, pp. 1–90.

558 Egli, M., Merkli, C., Sartori, G., Mirabella, A., Plötze, M., 2008. Weathering, mineralogical evolution
 559 and soil organic matter along a Holocene soil toposequence on carbonate-rich materials.
 560 *Geomorphology* 97, 675–696.

561 Egli, M., Sartori, G., Mirabella, A., Favilli, F., Giaccai, D., Delbos, E., 2009. Effect of north and south
 562 exposure on organic matter in high Alpine soils. *Geoderma* 149, 124-136.

563 Egli, M., Mavris, C., Mirabella, A., Giaccai, D., Kägi, B., Haeberli, W., 2010. Soil organic matter
 564 formation along a chronosequence in the Morteratsch proglacial area (Upper Engadine,
 565 Switzerland). *Catena* 82, 61–69.

566 Egli, M., Favilli, F., Krebs, R., Pichler, B., Dahms, D., 2012. Soil organic carbon and nitrogen
 567 accumulation rates in cold and alpine environments over 1 Ma. *Geoderma* 183-184.

568 Eskelinen, A., Stark, S., Mannisto, M., 2009. Links between plant community composition, soil
 569 organic matter quality and microbial communities in contrasting tundra habitats. *Oecologia* 161,
 570 113–123.

571 Eusterhues, K., Rumpel, C., Kögel-Knabner, I., 2005. Stabilization of soil organic matter isolated via
572 oxidative degradation. *Organic Geochemistry* 36, 1567-1575.

573 Favilli, F., Egli, M., Cherubini, P., Sartori, G., Haeberli, W., Delbos, E., 2008. Comparison of different
574 methods of obtaining a resilient organic matter fraction in Alpine soils. *Geoderma* 145, 355-369.

575 Favilli, F., Egli, M., Brandová, D., Ivy-Ochs, S., Kubik, P., Cherubini, P., Mirabella, Al., Sartori, G.,
576 Giaccari, D., Haeberli, W., 2009. Combined use of relative and absolute dating techniques for
577 detecting signals of Alpine landscape evolution during the late Pleistocene and early Holocene.
578 *Geomorphology* 112, 48-66.

579 Favilli, F., Cherubini, P., Collenberg, M., Egli, M., Sartori, G., Schoch, W., Haeberli, W., 2010.
580 Charcoal fragments of Alpine soils as an indicator of landscape evolution during the Holocene in
581 Val di Sole (Trentino, Italy). *The Holocene* 20, 67-79.

582 FOEN (Federal Office for the Environment), 2005. Hinweiskarte der potentiellen
583 Permafrostverbreitung in der Schweiz, Blatt Bergün 2005 (Map of the potential permafrost
584 distribution of Switzerland, map sheet Bergün 2005). Bundesamt für Umwelt BAFU, Bern.

585 Frauenfelder, R., Haeberli, W., Hoelzle, M., Maisch, M., 2001. Using relict rockglaciers in GIS-based
586 modelling to reconstruct Younger Dryas permafrost distribution patterns in the Err-Julier area, Swiss
587 Alps. *Norsk Geografisk Tidsskrift* 55, 195-202.

588 Gleixner, G., Czimczik, C.J., Kramer, C., Luehker, B., Schmidt, M.W.I., 2001. Plant compounds and
589 their turnover and stabilization as soil organic matter. In: Schulze, E.D., et al. (Ed.), *Global*
590 *Biogeochemical Cycles in the Climate System*. Academic Press, San Diego, 201–215.

591 Guo, Y., Bustin, R.M., 1997. Micro-FTIR spectroscopy of liptinite macerals in coal. *International*
592 *Journal of Coal Geology* 36, 259-275.

593 Guo, L.B., Gifford, R.M., 2002. Soil carbon stocks and land use change: a meta analysis. *Global*
594 *Change Biology* 8, 345-360.

595 Hagedorn, F., Moeri, A., Walthert, L., Zimmermann, S., 2010. Kohlenstoff in Schweizer Waldböden –
 596 bei Klimaerwärmung eine potenzielle CO₂-Quelle. Schweizerische Zeitschrift für Forstwesen 161,
 597 530-535.

598 Haugland, J.E., Beatty, S.W., 2005. Vegetation establishment, succession and microsite frost
 599 disturbances on glacier forelands within patterned ground chronosequences. Journal of
 600 Biogeography 32, 145–153.

601 Helfrich, M., Ludwig, B., Buurman, P., Flessa, H., 2006. Effect of land use on the composition of soil
 602 organic matter in density and aggregate fractions as revealed by solid-state ¹³C NMR spectroscopy.
 603 Geoderma 136, 331–341.

604 Helfrich, M., Flessa, H., Mikutta, R., Dreves, A., Ludwig, B., 2007. Comparison of chemical
 605 fractionations methods for isolating stable soil organic carbon pools. European Journal of Soil
 606 Science 58, 1316-1329.

607 Hitz, C., Egli, M., Fitze, P., 2001. Below-ground and above-ground production of vegetational organic
 608 matter along a climosequence in alpine grasslands. Journal of Plant Nutrition and Soil Science 164,
 609 389–397.

610 Hitz, C., Egli, M., Fitze, P., 2002. Determination of the sampling volume for representative analysis of
 611 alpine soils. Zeitschrift für Pflanzenernährung und Bodenkunde 165, 326-331.

612 Hugelius, G., Kuhry, P., 2009. Landscape partitioning and environmental gradient analyses of soil
 613 organic carbon in a permafrost environment. Global Biogeochemical Cycles 23, GB3006.

614 Jacot, K.A., Lüscher, A., Nosberger, J., Hartwig, U.A., 2000. Symbiotic N-2 fixation of various legume
 615 species along an altitudinal gradient in the Swiss Alps. Soil Biology and Biochemistry 32, 1043–
 616 1052.

617 Jackson, R.B., Banner, J.L., Jobbagy, E.G., Pockman, W.T., Wall, D.H., 2002. Ecosystem carbon loss
 618 with woody plant invasion of grasslands. Nature 418, 623–626.

619 Kääb, A., Chiarle, M., Raup, B., Schneider, C., 2007. Climate change impacts on mountain glaciers and
620 permafrost. *Global and Planetary Change* 56, vii–ix.

621 Kaiser, C., Meyer, H., Biasi, C., Rusalimova, O., Barsukov, P., Richter, A., 2007. Conservation of soil
622 organic matter through cryoturbation in arctic soils in Siberia. *Journal of Geophysical Research-*
623 *Biogeosciences* 112, G02017.

624 Kleber, M., Sollins, P., Sutton, R., 2007. A conceptual model of organo-mineral interactions in soils:
625 self-assembly of organic molecular fragments into zonal structures on mineral surfaces.
626 *Biogeochemistry* 85, 9–24.

627 Kneisel, C., Hauck, C., Vonder Mühll, D., 2000. Permafrost below the timberline confirmed and
628 characterized by geoelectrical resistivity measurements, Bever Valley, eastern Swiss Alps.
629 *Permafrost and Periglacial Processes* 11, 295-304.

630 Kneisel, C., Hauck, C., 2003. Multi-method geophysical investigation of a sporadic permafrost
631 occurrence. In: Schrott, L. Hoerd, A., Dikau, R. (eds.), *Geophysical methods in geomorphology*.
632 *Zeitschrift für Geomorphologie Suppl.* 132, 145-159.

633 Kneisel, C., Schwindt, D. 2008. Geophysical mapping of isolated permafrost lenses at a sporadic
634 permafrost site at low altitude in the Swiss Alps. *Proceedings of the Ninth International Conference*
635 *on Permafrost (Fairbanks, Alaska)* , 959-964.

636 Kögel-Knabner, I., 2002. The macromolecular organic composition of plant and microbial residues as
637 inputs to soil organic matter. *Soil Biology and Biochemistry* 34, 139-162.

638 Kögel-Knabner, I., Guggenberger, G., Kleber, M., Kandeler, E., Kalbitz, K., Scheu, S., Eusterhues, K.,
639 Leinweber, P., 2008. Organo-mineral associations in temperate soils: integrating biology,
640 mineralogy, and organic matter chemistry. *Journal of Plant Nutrition and Soil Science* 171, 61–82.

641 Kottek, M., Grieser, J., Beck, C., Rudolf, B., Rubel, F., 2006. World Map of Köppen-Geiger Climate
642 classification updated. *Meteorologische Zeitschrift* 15, 259-263.

643 Krull, E.S., Baldock, J.A., Skjemstad, J.O., 2003. Importance of mechanisms and processes of the
 644 stabilisation of soil organic matter for modelling carbon turnover. *Functional Plant Biology* 30, 207–
 645 222.

646 Leifeld, J., Kögel-Knabner, I., 2005. Soil organic matter fractions as early indicators for carbon stock
 647 changes under different land-use? *Geoderma* 124, 143-155.

648 Leifeld, J., Zimmermann, M., Fuhrer, J., Conen, F., 2009. Storage and turnover of carbon in grassland
 649 soils along an elevation gradient in the Swiss Alps. *Global Change Biology* 15, 668–679.

650 Leifeld, J., Bassin, S., Conen, F., Hajdas, I., Egli, M., Fuhrer, J., 2012. Control of soil pH on turnover
 651 of belowground organic matter fractions in subalpine grasslands. *Biogeochemistry*, DOI
 652 10.1007/s10533-011-9689-5.

653 Maisch, M., Wipf, A., Denzler, B., Battaglia, J., Benz, C., 1999. Die Gletscher der Schweizer Alpen.
 654 Gletscherhochstand 1850, aktuelle Vergletscherung, Gletscherschwund- Szenarien. Schlussbericht
 655 NFP 31 Projekt, vdf-Hochschulverlag ETH Zürich.

656 Mikutta, R., Kleber, M., Torn, M.S., Jahn R., 2006. Stabilization of organic matter: association with
 657 minerals or chemical recalcitrance? *Biogeochemistry* 77, 25-56.

658 Nötzli, J., Gruber, S., 2005. Alpiner Permafrost – Ein Überblick. *Jahrbuch des Vereins zum Schutz der*
 659 *Bergwelt* 70, 111-121.

660 Ohtonen, R., Fritze, H., Pennanen, T., Jumpponen, A., Trappe, J., 1999. Ecosystem properties and
 661 microbial community changes in primary succession on a glacier forefront. *Oecologia* 119, 239–
 662 246.

663 Ozenda, P., Borel, J.L., 1991. Les conséquences écologiques possibles des changements climatiques
 664 dans l'Arc alpin. *Rapport Futuralp*, vol. 1. ICALPE, Le Bourget-du-Lac, France.

665 Piccolo, A., Mirabella, A., 1985. Effetto di differenti estraenti inorganici e organici su alcune proprietà
 666 chimiche e chimico-fisiche di un acido umico. *Annali dell'Istituto Sperimentale per lo Studio e*
 667 *Difesa Suolo* XVI, 159-168.

668 Plante, A.F., Chenu, C., Balabane, M., Mariotti, A., Righi, D., 2004. Peroxide oxidation of clay-
 669 associated organic matter in a cultivation chronosequence. *European Journal of Soil Science* 55,
 670 471-478.

671 Reimer, P.J., Baillie, M.G.L., Bard, E., Bayliss, A., Beck, J.W., Blackwell, P.G., Bronk Ramsey, C.,
 672 Buck, C.E., Burr, G.S., Edwards, R.L., Friedrich, M., Grootes, P.M., Guilderson, T.P., Hajdas, I.,
 673 Heaton, T.J., Hogg, A.G., Hughen, K.A., Kaiser, K.F., Kromer, B., McCormac, F.G., Manning,
 674 S.W., Reimer, R.W., Richards, D.A., Southon, J.R., Talamo, S., Turney, C.S.M., van der Plicht, J.,
 675 Weyhenmeyer, C.E., 2009. INTCAL09 and MARINE09 radiocarbon age calibration curves, 0-
 676 50,000 years cal BP. *Radiocarbon* 51, 1111-1150.

677 Sanderman, J., Amundson, R.G., Baldocchi, D.D., 2003. Application of eddy covariance measurements
 678 to the temperature dependence of soil organic matter mean residence time. *Global Biogeochemical*
 679 *Cycles* 17, 1061.

680 Schimel, D.S., Braswell, B.H., Holland, E.A., McKeown, R., Ojima, D.S., Painter, T.H., Parton, W.J.,
 681 Townsend, A.R., 1994. Climatic, edaphic, and biotic controls over storage and turnover of carbon in
 682 soils. *Global Biogeochemical Cycles* 8, 279–293.

683 Schmidt, M.W.I., Torn, M.S., Abiven, S., Dittmar, T., Guggenberger, G., Janssens, I.A., Kleber, M.,
 684 Kögel-Knabner, I., Lehmann, J., Manning, D.A.C., Nannipieri, P., Rasse, D.P., Weiner, S.,
 685 Trumbore, S.E., 2011. Persistence of soil organic matter as an ecosystem property. *Nature* 478, 49-
 686 56.

687 Schuur, E.A.G., Bockheim, J., Canadell, J.G., Euskirchen, E., Field, C.B., Goryachkin, S.V.,
 688 Hagemann, S., Kuhry, P., Lafleur, P.M., Lee, H., Mazhitova, G., Nelson, F.E., Rinke, A.,
 689 Romanovsky, V.E., Shiklomanov, N., Tarnocai, C., Veevsky, S., Vogel, J.G., Zimov, S.A., 2008.
 690 Vulnerability of permafrost carbon to climate change: Implications for the global carbon cycle.
 691 *BioScience* 58, 701-714.

692 Schwarb, M., Daly, C., Frei, C., Schär, C., 2000. Mittlere jährliche Niederschlagshöhe im
 693 europäischen Alpenraum 1971-1990. In Hydrologischer Atlas der Schweiz, Blatt 2.6.

694 Seeber, J., Langel, R., Meyer, E., Traugott, M., 2009. Dwarf shrub litter as a food source for macro-
 695 decomposers in alpine pastureland. *Applied Soil Ecology* 41,178–184.

696 Senesi, N., V. D’Orazio, G. Ricca, 2003. Humic acids in the first generation of EUROSOLS.
 697 *Geoderma* 116, 325-344.

698 Soil Survey Staff, 2010. Keys to Soil Taxonomy, 11th edition. USDA (United States Department of
 699 Agriculture), NRCS (National Resources Conservation Service), Washington, DC.

700 Sollins, P., Swanston, C., Kleber, M., Filley, T., Kramer, M., Crow, S., Caldwell, B.A., Lajtha, K.,
 701 Bowden, R., 2006. Organic C and N stabilization in a forest soil: Evidence from sequential density
 702 fractionation. *Soil Biology and Biochemistry* 38, 3313–3324.

703 Sollins, P., Kramer, M.G., Swanston, C., Lajtha, K., Filley, T., Aufdenkampe, A.K., Wagai, R.,
 704 Bowden, R.D., 2009. Sequential density fractionation across soils of contrasting mineralogy:
 705 evidence for both microbial- and mineral-controlled soil organic matter stabilization.
 706 *Biogeochemistry* 96, 209–231.

707 Stevenson, F.J., 1994. Humus chemistry: genesis, composition, reactions. 2nd ed., Wiley, New York.

708 Suter, J., 1981. Gletschergeschichte des Oberengadins: Untersuchungen von Gletscherschwankungen
 709 in der Err-Julier-Gruppe. Ph.D. thesis, Univ. Zürich, Physische Geographie, vol. 2.

710 Tan, K. H., 2003. Humic Matter in Soil and the Environment. Marcel Dekker, Inc., New York.

711 Tarnocai, C., 2000. Carbon pools in soils of the arctic, subarctic, and boreal regions of Canada. In: Lal,
 712 R., Kimble, J.M., Stewart, B.A. (Eds.), *Global Climate Change and Cold Regions Ecosystems*.
 713 *Advances in Soil Science*, Lewis Publishers, Boca Raton, 91–103.

714 Tarnocai, C., Canadell, J.G., Schuur, E.A.G., Kuhry, P., Mazhitova, G., Zimov, S., 2009. Soil organic
 715 carbon pools in the northern circumpolar permafrost region. *Global Biogeochemical Cycles* 23,
 716 GB2023.

717 Tate, K.R., Scott, N.A., Ross, D.J., Parshotam, A., Claydon, J.J., 2000. Plant effects on soil carbon
 718 storage and turnover in a montane beech (*Nothofagus*) forest and adjacent tussock grassland in New
 719 Zealand. *Australian Journal of Soil Research* 38, 685-835.

720 Ugolini, F.C., Corti, G., Certini, G., 2006. Pedogenesis in the sorted patterned ground of Devon
 721 plateau, Devon Island, Nunavut, Canada. *Geoderma* 136, 87-106.

722 Von Lützow, M., Kögel-Knabner, I., Ekschmitt, K., Matzner, E., Guggenberger, G., Marschner, B.,
 723 Flessa, H., 2006. Stabilization of organic matter in temperate soils: mechanisms and their relevance
 724 under different soil conditions – a review. *European Journal of Soil Science* 57, 426–445.

725 Von Lützow, M., Kögel-Knabner, I., 2009. Temperature sensitivity of soil organic matter
 726 decomposition-what do we know? *Biology and Fertility of Soils* 46, 1-15.

727 WRB, 2006. World Reference Base for Soil Resources (2nd ed.). World Soil Resources Reports No.
 728 103, FAO, Rome.

729 Zimov, S.A., Schuur, E.A.G., Chapin III, F.S., 2006. Permafrost and the global carbon budget. *Science*
 730 312, 1612-1613.

Table 1. Chemical composition of the skeleton fraction of the parent material (‘Julier Granite’; Böhlert et al., 2011a).

	SiO ₂ g/kg	Al ₂ O ₃ g/kg	Fe ₂ O ₃ g/kg	CaO g/kg	MgO g/kg	K ₂ O g/kg	Na ₂ O g/kg	MnO g/kg	P ₂ O ₅ g/kg	TiO ₂ g/kg	ZrO ₂ g/kg	LOI g/kg
Mean	686.3	148.8	30.9	22.9	9.6	28.8	29.8	0.60	0.98	4.09	0.20	18.3
SD	± 18.5	± 5.1	± 5.4	± 14.4	± 2.5	± 5.5	± 1.9	± 0.11	± 0.15	± 0.60	± 0.03	± 4.1

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Table 2. Characteristics of the study sites.

[illegible]

B1m	46°32'9°47'	2674	30	0	-2.6	>1050	Granite/Gneiss (Julier Granite)	Alpine tundra (<i>Caricetum curvulae</i>)	Haplic Cambisol (Turbic)
B2m	46°32'9°48'	2648	30	36	-2.6	>1050	Granite/Gneiss (Julier Granite)	Alpine tundra (<i>Caricetum curvulae</i>)	Haplic Cambisol (Turbic)
B3m	46°32'9°48'	2663	30	30	-2.6	>1050	Granite/Gneiss (Julier Granite)	Alpine tundra (<i>Caricetum curvulae</i>)	Haplic Cambisol (Turbic)
Albula A1m	46°34'9°53'	2620	70	62	-2.6	>1050	Granite/Gneiss (Julier Granite)	Alpine tundra (<i>Caricetum curvulae</i>)	Entic Podzol
A2m	46°34'9°53'	2620	55	62	-2.6	>1050	Granite/Gneiss (Julier Granite)	Alpine tundra (<i>Caricetum curvulae</i>)	Entic Podzol
A3m	46°34'9°53'	2616	50	70	-2.6	>1050	Granite/Gneiss (Julier Granite)	Alpine tundra (<i>Caricetum curvulae</i>)	Entic Podzol
<i>Permafrost sites: Subalpine</i>									
Spinas S1m	46°26'9°57'	1820	20	57	1.0	1050	Granite rich slope deposits	Natural forest (<i>Larici-Pinetum cembrae</i>)	Haplic Cambisol (Dystric)
S2m	46°26'9°57'	1820	20	57	1.0	1050	Granite rich slope deposits	Natural forest (<i>Larici-Pinetum cembrae</i>)	Haplic Cambisol (Dystric)
S3m	46°33'9°51'	1808	40	46	1.0	1050	Granite rich slope deposits	Natural forest (<i>Larici-Pinetum cembrae</i>)	Haplic Cambisol (Dystric)
S4m	46°33'9°51'	1809	40	53	1.0	1050	Granite rich slope deposits	Natural forest (<i>Larici-Pinetum cembrae</i>)	Haplic Podzol
Wm2, Wm3, Wm4, Wm7, Wm9	46°33'9°51'	1719	40	46	1.0	1050	Granite rich slope deposits	Natural forest (<i>Larici-Pinetum cembrae</i>)	-

¹World geodetic system (WGS84)

²MAAT = mean annual air temperature, MAP = mean annual precipitation (according to EDI, 1992; Leidlmaier, 1996; Schwarb et al., 2000).

³WRB (IUSS Working Group 2006)

Table 3. Major IR absorption bands and assignments (Piccolo and Mirabella, 1985; Stevenson, 1994; Guo and Bustin, 1997; Senesi et al., 2003; Tan, 2003).

Band	Wave number cm ⁻¹	Assignment
1	2980-2880	Aliphatic C-H stretching (aliphatic methyl and methylene groups)
2	1725-1710	C=O stretching of COOH, aldehydes and ketones
3	1660-1630	C=O stretching of amide groups, quinone C=O and/or C=O of H-bonded conjugated ketones
4	1620-1600	Aromatic C=C, strongly H-bonded C=O of conjugated ketones
5	1535-1500	Aromatic rings, amide II vibrations
6	1495-1470	N-H stretching of proteic amides
7	1470-1440	Aliphatic C-H stretching
8	1413-1333	OH deformation and C-O stretching of phenolic groups
9	1190-1127	C-OH stretching of aliphatic, alcoholic O-H
10	1116-1050	Secondary alcohols
11	1080-1030	C-O stretching of polysaccharide

Table 4
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Table 4. Physical and chemical characteristics including labile and stable organic carbon fractions (H₂O₂ treated samples) of the fine earth (<2mm).

Site	Horizon	Depth (cm)	Skeleton (wt.%)	Bulk density (g cm ⁻³)	Sand (g kg ⁻¹)	Silt (g kg ⁻¹)	Clay (g kg ⁻¹)	pH (CaCl ₂)	C _{tot} (g kg ⁻¹)	N _{tot} (g kg ⁻¹)	C _{tot} / N _{tot}	C _{stable} (g kg ⁻¹)	N _{stable} (g kg ⁻¹)	C _{stable} / N _{stable}	C _{labile} (g kg ⁻¹)	N _{labile} (g kg ⁻¹)	C _{labile} /N _{labil}	C _{rec} ¹⁾ %	N _{rec} ¹⁾ %
Non-permafrost sites: Alpine																			
Bever																			
B1o	O	0-10	2	0.4	122	577	301	4.0	197	15	13	7.9	0.9	9	189.1	14.1	13	4.0	6.0
	OE	10-30	35	-	565	197	238	4.0	85.4	7.4	12	2.0	0.3	6	83.4	7.1	12	2.3	4.1
	Bs	30-50	54	1.25	760	214	26	4.3	25.9	3.4	8	2.7	2.7	1	23.2	0.6	38	10.4	79.4
	C	50-75	77	1.54	875	111	14	4.4	18.2	2.4	8	2.1	0	-	16.1	2.4	7	11.5	-
B2o	A	0-10	22	0.77	-	-	-	4.1	99.3	7.7	13	3.1	0.3	11	96.2	7.4	13	3.1	3.9
	AC	10-30	52	1.1	-	-	-	4.1	63.5	6.2	10	1.9	0.2	11	61.6	6.0	10	3.0	3.2
	bA	30-50	50	1.1	-	-	-	4.3	47.8	5.0	10	3.7	2.5	2	44.2	2.5	18	7.7	50.0
	R	>50	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
B3o	O	0-10	36	0.54	-	-	-	3.9	110.0	7.8	14	4.2	0.6	7	105.8	7.2	15	3.8	7.7
	AE	10-30	55	1.2	-	-	-	4.1	107.0	7.1	15	4.7	0.3	19	102.3	6.9	15	4.4	4.2
	Bhs	30-50	54	1.2	-	-	-	4.3	42.2	3.6	12	3.1	2.8	1	39.1	0.8	50	7.3	77.8
	Bs	50-60	49	1.37	-	-	-	4.5	13.7	2.3	6	2.2	0	-	11.5	2.3	5	16.1	-
	C	60-70	65	-	-	-	-	4.6	6.1	1.4	4	2.1	0	-	4	1.4	3	34.4	-
Albula																			
A1o	AE	0-10	19	0.75	483	341	176	3.8	33.0	3.0	11	2.5	1.9	1	30.5	1.1	28	7.6	63.3
	BA	10-20	32	1.1	448	371	180	4.1	15.9	2.3	7	2.6	2.1	1	13.3	0.3	53	16.4	91.3
	Bhs1	20-40	49	1.1	485	345	170	4.1	15.0	2.1	7	2.5	1.5	2	12.5	0.5	23	16.7	71.4
	Bhs2	40-60	60	1.1	723	224	52	4.4	27.9	1.9	15	3.4	0	-	24.5	1.9	13	12.2	-
	C	60-70	49	-	653	305	42	4.5	10.9	0.2	50	2.5	0	-	8.4	0.2	38	22.9	-
A2o	AE	0-10	50	0.74	-	-	-	3.8	73.9	5.5	13	3.6	2.6	1	70.3	2.9	24	4.9	47.3
	Bhs1	10-20	52	1	-	-	-	4.4	39.8	4.1	10	2.6	2.1	1	37.2	2.0	19	6.5	51.2
	Bhs2	20-35	56	1	-	-	-	4.2	14.8	2.8	5	2.3	1.8	1	12.5	1.0	12	15.5	64.3
	C	35-45	68	-	-	-	-	4.5	8.7	1.6	5	2.9	0	-	5.8	1.6	4	33.3	-
A3o	A	0-10	17	0.85	-	-	-	3.5	79.6	7.4	11	3.5	2.6	1	76.1	4.8	16	4.4	35.1
	AE	10-20	46	0.85	-	-	-	3.8	43.6	5.1	9	3.9	2.1	2	39.7	3.0	13	8.9	41.2

	Bh1	20-40	65	1.07	-	-	-	4.2	32.3	3.6	9	2.7	2.1	1	29.6	1.6	19	8.4	58.3
	Bh2	40-60	62	1.07	-	-	-	4.3	29.2	3.1	9	2.8	3	1	26.4	0.2	168	9.6	96.8
	C	60-75	60	-	-	-	-	4.5	13.5	2.5	5	2.4	0	-	11	2.5	5	17.8	-
<i>Non-permafrost sites: Subalpine</i>																			
Spinas																			
S1o	E	0-10	69	1.19	728	205	67	4.0	58.4	3.2	18	5.6	1.0	6	52.7	2.3	23	9.6	31.3
	Bs1	10-20	57	-	750	214	36	3.4	49.9	3.8	13	5.2	0.7	7	44.7	3.1	15	10.4	18.4
	Bs2	20-30	68	1.18	749	220	31	4.3	37.7	2.2	18	5.0	1.3	4	32.6	0.8	39	13.3	59.1
	Bs3	30-40	60	-	778	184	38	4.4	35.1	2.7	13	3.5	0	-	31.6	2.7	12	10.0	-
	B	50-60	58	1.13	649	233	118	4.4	31.9	2.7	12	3.8	2.6	2	28.1	0.2	166	11.9	96.3
	R	>60	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
S2o	E	0-10	36	-	-	-	-	3.8	63.5	5.8	11	3.7	1.0	4	59.8	4.8	12	5.8	17.2
	Bs1	10-20	40	-	-	-	-	4.2	31.0	3.4	9	3.1	0.8	4	27.9	2.7	11	10.0	23.5
	Bs2	40-50	54	-	-	-	-	4.5	19.8	2.3	9	2.9	2.3	1	16.9	0	0	14.6	100.0
	B	70-80	82	-	-	-	-	4.4	15.8	2.4	7	2.5	0	-	13.3	2.4	6	15.8	-
	R	>80	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
S3o	A	0-10	63	0.51	-	-	-	3.8	65.7	3.2	21	7.4	1.0	8	58.3	2.2	26	11.3	31.3
	E	10-20	54	0.51	-	-	-	3.8	91.5	3.7	25	9.4	0.7	13	82.1	3.0	28	10.3	18.9
	Bh1	20-30	83	0.93	-	-	-	3.7	151.5	4.5	34	11.0	1.9	6	140.5	2.6	55	7.3	42.2
	Bh2	30-40	59	0.93	-	-	-	3.7	130.0	4.1	32	8.2	2.0	4	121.8	2.1	58	6.3	48.8
	IIE	40-50	41	0.79	-	-	-	3.7	25.3	1.2	21	3.0	0	-	22.3	1.2	19	11.9	-
	Bs	60-100	67	-	-	-	-	4.6	29.2	0.7	42	3.8	0	-	25.4	0.7	36	13.0	-
	R	>100	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Wo5	A	0-15	28	0.55	-	-	-		71.7	6.6	11	5.6	2.1	3	66	4.5	15	7.8	31.8
	B	15-30	50	1.22	-	-	-		19.3	2.6	7	2.5	0	-	16.8	2.6	6	13.0	-
Wo6	A	0-15	38	0.55	-	-	-		81	5.9	14	5.2	1.9	3	75.8	4.0	19	6.4	32.2
	B	15-30	27	1.22	-	-	-		47.1	3.6	13	3.6	1.9	2	43.4	1.7	26	7.6	52.8
Wo8	A	0-15	48	0.55	-	-	-		31.6	4.1	8	3.3	2.0	2	28.3	2.1	13	10.4	48.8
	B	15-30	48	1.22	-	-	-		22.8	3.2	7	3.2	1.9	2	19.6	1.3	15	14.0	59.4

Wo9	A	0-15	18	0.55	-	-	-		146.5	9.0	16	9.0	1.4	7	137.5	7.6	18	6.1	15.6
	B	15-30	21	1.22	-	-	-		52.1	4.5	12	2.4	2.5	1	49.7	2.0	25	4.6	55.6
Wo10	A	0-10	40	0.55	-	-	-		128.0	8.5	15	11.2	1.8	6	116.8	6.7	18	8.8	21.2
	B	15-20	33	1.22	-	-	-		37.8	3.9	10	3.1	1.0	6	34.8	3.4	10	8.2	25.6
<i>Permafrost sites: Alpine</i>																			
Bever																			
B1m	A	0-10	38	1.56	793	130	77	4.5	15.8	3.1	5	1.2	0.1	17	14.6	3.0	5	7.6	3.2
	Bw1	10-30	39	1.56	848	89	63	4.5	12.8	2.2	6	0.9	0.1	13	11.9	2.2	6	7.0	4.5
	Bw2	30-50	40	1.57	786	133	81	4.5	20.1	2.6	8	2.4	0	-	17.7	2.6	7	11.9	-
	BC	50-60	52	1.67	814	103	83	4.6	11.6	2.1	6	2.3	0	-	9.3	2.1	4	19.8	-
	R	>60	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
B2m	Ah	0-10	31	1.34	-	-	-	4.6	19.6	3.1	6	1.2	0.1	13	18.4	3.0	6	6.1	3.2
	Bw1	10-30	26	1.34	-	-	-	4.6	12.1	2.6	5	2.0	0.1	35	10.0	2.5	4	16.5	3.8
	Bw2	30-35	18	1.36	-	-	-	4.6	12.6	1.9	7	2.6	0	-	10.0	1.9	5	20.6	-
	Bw3	35-50	26	1.33	-	-	-	4.6	14.9	2.5	6	2.4	0	-	12.5	2.5	5	16.1	-
	C	50-80	32	1.74	-	-	-	4.6	12.7	2.3	5	2.2	0	-	10.5	2.3	4	17.3	-
B3m	Ah	0-10	29	1.11	-	-	-	4.4	29.6	3.8	8	2.5	1.4	2	27	2.4	11	8.4	36.8
	Bw	10-30	35	1.27	-	-	-	4.6	11.5	2.7	4	2.4	1.9	1	9	0.8	11	20.9	70.4
	BC1	30-50	42	1.61	-	-	-	4.7	7.2	1.7	4	2.0	0	-	5.2	1.7	3	27.8	-
	BC2	50-60	43	1.61	-	-	-	4.7	6.1	2.0	3	1.6	0	-	4.5	2.0	2	26.2	-
	R	>60	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Albula																			
A1m	A1	0-10	31	1.15	543	293	164	3.9	27.4	2.3	12	2.5	1.7	2	24.9	0.6	45	9.1	73.9
	A2	10-20	34	1.29	512	332	156	4.0	20.8	1.8	12	2.9	1.7	2	17.9	0.2	118	13.9	94.4
	AE	20-40	32	1.18	524	313	163	4.1	21.4	1.3	16	2.7	0	-	18.7	1.3	14	12.6	-
	Bs	40-80	54	1.42	619	272	109	4.6	22.1	1.2	18	3.3	0	-	18.8	1.2	16	14.9	-
	BC	80-100	67	1.62	674	277	49	4.3	6.6	0.3	26	2.4	0	-	4.1	0.3	17	36.4	-
	R	>100	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
A2m	AE1	0-10	48	0.78	-	-	-	3.6	71.2	4.1	18	2.8	2.4	1	68.4	1.7	41	3.9	58.5
	AE2	10-20	68	0.93	-	-	-	3.9	42.2	3.8	11	2.8	1.4	2	39.3	2.5	16	6.6	36.8
	Bhs	20-45	55	1.33	-	-	-	4.3	20.4	1.7	12	2.3	1.4	2	18.1	0.4	49	11.3	82.4

	B	45-80	62	1.57	-	-	-	4.5	10.3	2.4	4	2.4	0	-	7.9	2.4	3	23.3	-
	C	80-100	44	1.42	-	-	-	4.6	3.6	0.2	17	2.3	0	-	1.3	0.2	6	63.9	-
A3m	A	0-10	62	0.66	-	-	-	3.8	96.4	5.6	17	4.6	2.2	2	91.8	3.5	27	4.8	39.3
	AE1	10-20	63	0.66	-	-	-	3.9	63.6	3.8	17	3.1	2.2	1	60.5	1.6	39	4.9	57.9
	AE2	20-35	67	0.66	-	-	-	4.1	55.6	2.7	21	3.3	2.1	2	52.2	0.6	90	5.9	77.8
	Bhs	35-60	59	1.20	-	-	-	4.5	32.3	1.4	23	2.6	0	-	29.7	1.4	21	8.0	-
	C	60-70	65	1.66	-	-	-	4.6	9.8	0.2	42	2.2	0	-	7.6	0.2	33	22.4	-

Permafrost sites: Subalpine

Spinas

S1m	A	0-10	55	1.15	633	218	149	3.8	40.0	4.3	9	3.8	1.1	4	36.2	3.3	11	9.5	25.6
	Bw	25-35	73	1.2	611	239	150	4.4	21.3	2.5	9	2.6	1.7	2	18.7	0.8	22	12.2	68.0
	BC	50-60	51	1.39	622	234	144	4.5	21.3	2.4	9	2.6	0	-	18.7	2.4	8	12.2	-
	R	>60	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
S2m	A	0-10	50	-	-	-	-	3.9	64.4	3.5	18	3.3	1.2	3	61.1	2.4	26	5.1	34.3
	Bw	25-35	61	-	-	-	-	4.1	45.2	2.2	21	3.8	1.2	3	41.4	1.0	43	8.4	54.5
	BC	50-60	51	-	-	-	-	4.1	49.0	2.4	20	3.8	0	-	45.2	2.4	19	7.8	-
	R	>60	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
S3m	Ah	0-10	52	0.65	-	-	-	3.5	93.2	5.5	17	5.5	1.2	5	87.7	4.3	20	5.9	21.8
	Bw1	10-20	53	0.84	-	-	-	4.0	36.6	3.0	12	2.8	0.8	4	33.8	2.2	15	7.7	26.7
	Bw2	20-30	73	0.84	-	-	-	4.1	32.6	3.0	11	3.1	1.3	2	29.5	1.7	17	9.5	43.3
	BC1	30-40	75	0.84	-	-	-	4.2	33.2	2.8	12	2.9	2.8	1	30.3	0	0	8.7	100.0
	BC2	40-50	68	0.84	-	-	-	4.3	29.3	2.1	14	3.7	0	-	25.6	2.1	12	12.6	
	C	50-60	60	-	-	-	-	4.3	33.1	3	11	3.6	2.3	2	29.5	0.7	46	10.9	76.7
S4m	AE	0-10	53	0.48	-	-	-	3.9	57.4	2.1	27	5.2	1.1	5	52.2	1.0	53	9.1	52.4
	Bh	10-20	59	0.48	-	-	-	3.9	35.8	0.8	45	4.9	0	-	30.9	0.8	39	13.7	-
	Bhs1	20-30	54	0.82	-	-	-	4.2	34.3	1.1	31	4.0	0	-	30.3	1.1	28	11.7	-
	Bhs2	30-40	62	0.73	-	-	-	4.3	39.9	0.7	57	3.8	0	-	36.1	0.7	52	9.5	-
	Bhs3	40-50	54	0.73	-	-	-	4.3	33.0	0.2	165	4.1	0	-	28.9	0.2	145	12.4	-
	Bhs4	50-60	64	0.73	-	-	-	4.3	40.5	0.8	51	5.2	0.2	32	35.3	0.6	55	12.8	25.0
	Bs1	60-70	62	-	-	-	-	4.3	36.8	0.8	46	3.7	0.2	23	33.1	0.6	52	10.1	25.0
	Bs2	70-90	73	-	-	-	-	4.3	20.9	0.6	35	2.6	0.2	14	18.3	0.4	44	12.4	33.3

	R	>90	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Wm2	A	0-10	16	0.16	-	-	-	200.0	11.2	18	18.2	1.3	14	181.9	10.0	18	9.1	11.6
	B	10-20	35	0.54	-	-	-	78.0	4.8	16	7.6	1.8	4	70.4	3.0	24	9.7	37.5
Wm3	A	0-15	34	0.16	-	-	-	280.5	14.0	20	19.2	1.3	15	261.4	12.7	21	6.8	9.3
	B	15-30	19	0.54	-	-	-	117.0	7.3	16	9.7	2.0	5	107.4	5.4	20	8.3	27.4
Wm4	A	0-15	14	0.16	-	-	-	354.5	13.8	26	25.3	0.7	34	329.2	13.0	25	7.1	5.1
	B	15-25	30	0.54	-	-	-	246.0	11.0	23	15.6	1.3	12	230.4	9.6	24	6.3	11.8
Wm7	A	0-15	25	0.16	-	-	-	146.0	7.8	19	6.7	1.9	4	139.3	5.9	24	4.6	24.4
	B	15-30	27	0.54	-	-	-	123.0	7.5	17	8.9	1.7	5	114.1	5.7	20	7.2	22.7
Wm9	A	0-15	38	0.16	-	-	-	195.5	11.2	17	12.6	1.4	9	182.9	9.8	19	6.4	12.5
	B	15-25	63	0.54	-	-	-	34.7	3.6	10	2.6	1.8	1	32.0	1.7	18	7.5	50.0

¹⁾ Recovery values

Table 5
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Table 5. Radiocarbon dating (obtained using ¹⁴C) of the bulk soil, the labile and stable organic carbon fractions (H₂O₂ treated samples) of the alpine sites (permafrost and non-permafrost).

Site	Depth (cm)	bulk C				labile C-fraction			stable C-fraction			
		pMC	Uncalibrated ages (y uncal. BP)	$\delta^{13}\text{C}$ (‰)	Calibrated ages (y cal. BP)	pMC	Uncalibrated ages (y uncal. BP)	Calibrated ages (y cal. BP)	pMC	Uncalibrated ages (y uncal. BP)	$\delta^{13}\text{C}$ (‰)	Calibrated ages (y cal. BP)
					<i>2 sigma range</i>			<i>2 sigma range</i>				<i>2 sigma range</i>
<i>non-permafrost</i>												
B1o	0-10	91	675 ± 30	-25.3 ± 1.1	631 - 679	92	640	562 - 591	82	1570 ± 30	-26.4 ± 1.1	1394 – 1530
	10-30	89	920 ± 25	-21.6 ± 1.1	777 - 920	89	890	760 - 799	75	2305 ± 30	-30.2 ± 1.1	2305 – 2357
B2o	0-10		modern	-	modern	-	-	modern		875 ± 35	-28.5 ± 1.1	725 – 835
	10-30	92	595 ± 35	-26.0 ± 1.1	538 - 654	93	551	532 - 555	76	2135 ± 35	-25.8 ± 1.1	2000 – 2162
B3o	0-10	86	1170 ± 30	-26.8 ± 1.1	1049 - 1175	86	1117	979 - 1057	71	2645 ± 30	-23.7 ± 1.1	2733 – 2796
	10-30	79	1815 ± 30	-26.0 ± 1.1	1691 - 1825	80	1743	1613 - 1700	64	3555 ± 30	-29.4 ± 1.1	3817 – 3927
<i>permafrost</i>												
B1m	0-10	76	2170 ± 30	-26.2 ± 1.1	2106 - 2311	76	2109	2041 - 2130	69	2980 ± 35	-24.7 ± 1.1	3061 – 3268
	10-30	68	3015 ± 40	-19.8 ± 1.1	3078 - 3341	68	3003	3161 - 3256	67	3180 ± 40	-25.7 ± 1.1	3335 – 3479
B2m	0-10	93	485 ± 30	-30.3 ± 1.1	499 - 545	95	394	469 - 501	77	2025 ± 35	-33.1 ± 1.1	1892 – 2064
	10-30	73	2495 ± 35	-22.1 ± 1.1	2456 - 2736	81	1613	1509 - 1543	31	9465 ± 35	-33.6 ± 1.1	10584 - 10788
B3m	0-10	83	1450 ± 30	-26.1 ± 1.1	1299 - 1389	84	1335	1271 - 1294	70	2785 ± 30	-29.7 ± 1.1	2838 – 2957
	10-30	68	3000 ± 30	-22.9 ± 1.1	3078 - 3269	71	2704	2766 - 2845	59	4205 ± 30	-30.8 ± 1.1	4627 - 4763

Table 6
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Table 6. Radiocarbon dating (obtained using ¹⁴C) of the bulk soil, the labile and stable organic carbon fractions (H₂O₂ treated samples) of the subalpine sites (permafrost and non-permafrost).

Site	Depth (cm)	bulk C				labile C-fraction			stable C-fraction			
		pMC	Uncalibrated ages (y uncal. BP)	$\delta^{13}\text{C}$ (‰)	Calibrated ages (y cal. BP)	pMC	Uncalibrated ages (y uncal. BP)	Calibrated ages (y cal. BP)	pMC	Uncalibrated ages (y uncal. BP)	$\delta^{13}\text{C}$ (‰)	Calibrated ages (y cal. BP)
					<i>2 sigma range</i>			<i>2 sigma range</i>				<i>2 sigma range</i>
<i>non-permafrost</i>												
Wo5	0-15	94	430 ± 25	-28.1 ± 1.1	461 - 525	95	359	434 - 480	84	1315 ± 25	-26.8 ± 1.1	1228 – 1295
						-	-	-		-	-	-
Wo6	0-15	100	-35 ± 25	-27.9 ± 1.1	41 - 60	102	modern	modern	72	2590 ± 25	-29.7 ± 1.1	2711 – 2760
	15-30		55 ± 25	-27.8 ± 1.1	31 - 84	-	-	-		-	-	-
Wo8	0-15	106	-500 ± 25	-28.2 ± 1.1	46 - 55	110	modern	modern	66	3295 ± 25	-28.5 ± 1.1	3453 – 3579
	15-30		modern	-27.8 ± 1.1	modern	-	-	-		-	-	-
Wo9	0-15	106	-560 ± 25	-27.9 ± 1.1	46 - 55	107	modern	modern	90	775 ± 25	-30.2 ± 1.1	673 – 731
	15-30		modern	-30.4 ± 1.1	modern	-	-	-		-	-	-
<i>permafrost</i>												
Wm2	0-15	102	-235 ± 25	-27.2 ± 1.1	46 - 55	104	modern	modern	89	855 ± 25	-29.6 ± 1.1	695 – 796
	15-30		400 ± 30	-30.6 ± 1.1	428 - 514	-	-	-		-	-	-
Wm3	0-15	108	-655 ± 25	-27.2 ± 1.1	46 - 54	109	modern	modern	96	310 ± 25	-29.6 ± 1.1	348 – 459
	15-30		modern	-24.2 ± 1.1	modern	-	-	-		-	-	-
Wm4	0-15	104	-400 ± 25	-28.2 ± 1.1	46 - 54	106	modern	modern	84	1295 ± 25	-30.6 ± 1.1	1178 – 1287
	15-30		145 ± 25	-23.5 ± 1.1	185 - 224	-	-	-		-	-	-

Figure captions

Fig. 1. Above: Study sites at the Val Bever and the Albula pass with the approximate locations of the sampling sites: (1) Bever, alpine; (2) Albula, alpine; (3) Spinas, subalpine. Below: Photographs of the sampling sites and their characteristic soil profiles. (1) alpine site: ‘non-permafrost’ (left) and ‘permafrost’ (right). (3) subalpine site: no difference in the terrain between both sites and typical soil profiles; ‘non-permafrost’ (left) and ‘permafrost’ (right).

Fig. 2. A) Carbon recovery after the H_2O_2 treatment for permafrost and non-permafrost soils. Alpine sites (Bever, Albula): left half of the plot; Subalpine site (Spinas): right half of the plot. B) C/N ratios of the bulk soil (untreated) and stable (H_2O_2 -treated) C-fraction of permafrost and non-permafrost soils. Alpine sites (Bever, Albula): left half of the plot; subalpine site (Spinas): right half of the plot.

Fig. 3. Stocks of total, labile and stable (recovery after the H_2O_2 -treatment) organic carbon in permafrost and non-permafrost soils. Alpine sites (Bever, Albula): left; subalpine site (Spinas): right. Carbon stocks were calculated down to the C-horizon or rock surface.

Fig. 4. Box plots (median, 25th- and 75th-percentile values, min. and max. values) of the relative proportion of A) secondary alcohols, B) C-O stretching of polysaccharide, C) OH deformation and C-O stretching of phenolic groups and D) N-H stretching of proteic amides of the bulk soil (untreated) and stable (H_2O_2 -treated) C-fractions of permafrost and non-permafrost soils. Alpine sites (Bever, Albula): left; subalpine site (Spinas): right.

Fig. 5. Carbon concentration in the different density fractions in the topsoil (0-15cm) of the subalpine site (N=3). Fractions of oPOM and mOM were additionally treated with an ultrasonic energy of 50 J ml^{-1} .

Fig. 6. Comparison of the SOC age range of paired samples (permafrost vs. non-permafrost sites and corresponding soil horizons). A) bulk C, B) labile C, C) stable C, D) stable C including additional subalpine sites (in red: Egli et al., 2009). Age cold soils = age range of the permafrost soils (this investigation) and additional subalpine (north-facing) sites and corresponding soil horizons (Egli et al., 2009). Age less cold soils = age range of the non-permafrost soils (this investigation) and additional subalpine (south-facing) sites and corresponding soil horizons (Egli et al., 2009).

Figure 1
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Figure 2
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Figure 3
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Alpine sites

Subalpine site

Figure 5
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Figure 6
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