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# Quaternary Science Reviews

## An annual-resolution stable isotope record from Swiss subfossil pine trees growing in the Late Glacial --Manuscript Draft--

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<b>Abstract:</b>	<p>Previous studies have suggested that the Late Glacial period (LG; ~ 14,600 – 11,700 cal BP) was characterised by abrupt and extreme climate variability over the European sector of the North Atlantic. The limited number of precisely dated, high-resolution proxy records, however, restricts our understanding of climate dynamics through the LG. Here, we present the first annually-resolved tree-cellulose stable oxygen and carbon isotope chronology (<math>\delta^{18}\text{O}_{\text{tree}}</math>, <math>\delta^{13}\text{C}_{\text{tree}}</math>) covering the LG between ~14,050 – 12,795 cal BP, generated from a Swiss pine trees (<i>P. sylvestris</i>; 27 trees, 1255 years). Comparisons of <math>\delta^{18}\text{O}_{\text{tree}}</math> with regional lake and ice core <math>\delta^{18}\text{O}</math> records reveal that LG climatic changes over the North Atlantic (as recorded by Greenland Stadials and Inter-Stadials) were not all experienced to the same degree in the Swiss trees. Possible explanations include: (1) LG climate oscillations may be less extreme during the summer in Switzerland, (2) tree-ring <math>\delta^{18}\text{O}</math> may capture local precipitation and humidity changes and/or (3) decayed cellulose and various micro-site conditions may overprint large-scale temperature trends found in other <math>\delta^{18}\text{O}</math> records. Despite these challenges, our study emphasises the potential to investigate hydroclimate conditions using subfossil pine stable isotopes.</p>

## Quaternary Science Reviews Submission

## An annual-resolution stable isotope record from Swiss subfossil pine trees growing in the Late Glacial

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1 Abstract

2 Previous studies have suggested that the Late Glacial period (LG; ~ 14,600 – 11,700  
3 cal BP) was characterised by abrupt and extreme climate variability over the European  
4 sector of the North Atlantic. The limited number of precisely dated, high-resolution  
5 proxy records, however, restricts our understanding of climate dynamics through the  
6 LG. Here, we present the first annually-resolved tree-cellulose stable oxygen and  
7 carbon isotope chronology ( $\delta^{18}\text{O}_{\text{tree}}$ ,  $\delta^{13}\text{C}_{\text{tree}}$ ) covering the LG between ~14,050 –  
8 12,795 cal BP, generated from a Swiss pine trees (*P. sylvestris*; 27 trees, 1255 years).

9 Comparisons of  $\delta^{18}\text{O}_{\text{tree}}$  with regional lake and ice core  $\delta^{18}\text{O}$  records reveal that LG  
10 climatic changes over the North Atlantic (as recorded by Greenland Stadials and Inter-  
11 Stadials) were not all experienced to the same degree in the Swiss trees. Possible  
12 explanations include: (1) LG climate oscillations may be less extreme during the  
13 summer in Switzerland, (2) tree-ring  $\delta^{18}\text{O}$  may capture local precipitation and humidity  
14 changes and/or (3) decayed cellulose and various micro-site conditions may overprint  
15 large-scale temperature trends found in other  $\delta^{18}\text{O}$  records. Despite these challenges,  
16 our study emphasises the potential to investigate hydroclimate conditions using  
17 subfossil pine stable isotopes.

18 Keywords: Tree-ring cellulose,  $\delta^{13}\text{C}$ ,  $\delta^{18}\text{O}$ , Late Glacial, Switzerland, Climate



19 Highlights

- 20 - Millennial-length stable isotope chronologies of  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  developed from  
21 27 Swiss subfossil pine trees, covering the Late Glacial (~14,050 – 12,795 cal  
22 BP);
- 23 - Two periods of extreme  $\delta^{18}\text{O}_{\text{tree}}$  depletions parallel known North Atlantic ‘cool  
24 periods’ recorded in Greenland  $\delta^{18}\text{O}_{\text{NGRIP}}$  (GI-1c, GS-1), while another LG  
25 oscillation (GI-1b) is not clearly expressed in  $\delta^{18}\text{O}_{\text{tree}}$
- 26 - Trees with less extensive decay have significantly higher inter-isotope  
27 ( $\delta^{18}\text{O}_{\text{tree}}$ :  $\delta^{13}\text{C}_{\text{tree}}$ ) correlations, and decay seemed to impact the stability of  
28  $\delta^{13}\text{C}_{\text{tree}}$  more strongly over  $\delta^{18}\text{O}_{\text{tree}}$

## 29 **1. Introduction**

### 30 1.1 The Late Glacial & climate proxy records

31 The Late Glacial (LG; ~ 14,600 – 11,700 cal BP) in the North Atlantic (NA) region is  
32 characterised by extreme and abrupt climate change, including at least four prominent  
33 “cold periods”. In Greenland, these events are defined by ice core  $\delta^{18}\text{O}$  downturns, i.e.  
34 substages within the Greenland Stadials (GS) and Inter-Stadials (GI); GI-1d  
35 (Rasmussen et al. 2006), GI-1c (Rasmussen et al. 2014), GI-1b (Björk et al. 1998) and  
36 GS-1 (Mayle et al. 1999, Rasmussen et al. 2006, Steffensen et al. 2008). Greenland  
37 ice core  $\delta^{18}\text{O}$  records are regarded as key datasets for comparison of hydroclimate  
38 indices across the NA region (Figure 3) (Alley et al. 1993, Rasmussen et al. 2014,  
39 Steffensen et al. 2008), due to their unrivalled temporal resolution (annual in parts)  
40 and extent (tens of thousands of years) as well as its sensitivity to NA precipitation and  
41 temperature variability.

42 In central Europe,  $\delta^{18}\text{O}$  in carbonates from several lake sediments offer sequences  
43 that can be directly compared to Greenland and portray broadly consistent signals of  
44 inferred temperature change ( $\delta^{18}\text{O}$  drop) in the low frequency domain (e.g. Lauterbach  
45 et al., 2011, Schwander et al. 2000, Siegenthaler et al. 1984, van Raden et al. 2012,  
46 2013, Verbruggen et al. 2010, von Grafenstein et al. 2000). Pronounced excursions in  
47 lake carbonate  $\delta^{18}\text{O}$  (negative) and speleothem  $\delta^{18}\text{O}$  (positive) records have  
48 pinpointed similar cold periods across central Europe including Germany (von  
49 Grafenstein et al. 1999, Rach et al., 2014), Switzerland (Lotter et al. 1992, 1999, 2000,  
50 Verbruggen et al. 2010) and Austria (Lauterbach et al. 2011). The characterisation and  
51 drivers of these temperature fluctuations, however, remains highly uncertain (Rach et  
52 al. 2014, Rasmussen et al. 2014). The low frequency climate variability inherent in lake

53 records provides information about decadal-scale variability on local to regional  
54 scales. However, the inter-annual features of such events are indistinguishable.

55 In this respect, tree-rings offer an annually resolved archive not only in terms of dating  
56 like varved lake sediments, but also with respect to proxy information from parameters  
57 (e.g. ring width, maximum latewood density, stable isotope ratios) measured on  
58 sequences of individual tree rings. However, due to the relatively short lifespans of  
59 individual trees, it is challenging to maintain low frequency climate information within  
60 millennial records (Cook et al. 1995, Esper et al. 2002), as some records have  
61 revealed limited power at multi-centennial time scales (e.g. Saurer et al. 2012, Cook  
62 et al. 2019). Fortunately, it has been demonstrated that subfossil tree-ring  $\delta^{18}\text{O}$   
63 records can capture large-scale climate deterioration, revealed through increased  
64 variability and distinct negative excursions (Pauly et al. 2018).

## 65 1.2 Climate proxies in tree-ring stable isotopes

66 The growth increments (ring width), wood density and stable isotope ratios of  
67 numerous modern species in Europe have been correlated significantly to relevant  
68 climatic parameters (temperature, precipitation, humidity) in numerous studies (e.g.  
69 Saurer et al. 2012, Büntgen et al. 2011, Esper et al. 2014). Signal strengths were found  
70 depending on the local limiting factors. Nevertheless, tree-ring widths, density and  
71 stable isotopes have the capacity to trace inter-annual variability and extremes, which  
72 are characteristic of climate change (e.g. drought, wet periods, storms).

73  $\delta^{13}\text{C}$  of tree-ring cellulose ( $\delta^{13}\text{C}_{\text{tree}}$ ) is based on the  $\delta^{13}\text{C}$  of the atmospheric  $\text{CO}_2$   
74 source which contains no direct climatic signal. Its climate signature originates from  
75 fractionations during photosynthesis at the leaf or needle level (Farquhar et al., 1982).  
76 In contrast,  $\delta^{18}\text{O}$  of tree-ring cellulose ( $\delta^{18}\text{O}_{\text{tree}}$ ) is related to the  $\delta^{18}\text{O}$  of the

77 precipitation source via soil water.  $\delta^{18}\text{O}$  of soil water constitutes the  $\delta^{18}\text{O}$  input to the  
78 arboreal system and represents an average  $\delta^{18}\text{O}$  over several precipitation events  
79 modified by partial evaporation from the soil (depending on soil texture and porosity)  
80 and by a possible time lag, depending on rooting depth (Saurer et al. 2012).  $\delta^{18}\text{O}_{\text{tree}}$  is  
81 further dependent on two tree-internal processes: evaporative  $^{18}\text{O}$ -enrichment of leaf  
82 or needle water via transpiration, as well as biochemical fractionations and isotopic  
83 exchange of  $\delta^{18}\text{O}$  with trunk water during cellulose biosynthesis (e.g. Roden et al.,  
84 2000, Barbour, 2007, Kahmen et al. 2011, Treydte et al., 2014 and citations therein).

### 85 1.2.1 Climate signals in stable isotopes of modern Swiss tree-ring records

86 Modern tree-ring stable isotope chronologies in Switzerland have been found to have  
87 relatively strong correlations with a variety of summer (JJA) meteorological  
88 parameters. For example,  $\delta^{18}\text{O}_{\text{tree}}$  has been correlated with summer precipitation  
89 extremes (*various species*: Saurer et al. 2008), relative humidity and sunshine duration  
90 (*Pinus sylvestris*: Saurer et al. 1997, *Larix decidua*: Kress et al. 2010, *Pinus sylvestris*,  
91 *Picea abies*, *Fagus sylvatica*, *Fraxinus excelsior*: Saurer et al. 2012) and  $\delta^{13}\text{C}_{\text{tree}}$  with  
92 summer temperature (*various species*: Saurer et al. 2008) and drought conditions  
93 (*Larix decidua*: Kress et al. 2010). Synoptic patterns and associated indices (e.g.  
94 cyclone and anticyclone activity, air pressure) have also been reconstructed from  
95 Swiss tree-ring  $\delta^{18}\text{O}_{\text{tree}}$ , with stronger correlations revealed during extreme years or  
96 periods (Saurer et al. 2012). Despite the success in tree-ring isotopes reflecting  
97 climate parameters, the relationships have been found to be unstable through time  
98 (Kress et al. 2010, Reynolds-Henne et al. 2007, Seftigen et al. 2011) as  
99 interdependency between meteorological variables also often changes depending on  
100 the climate state. As a result, Saurer et al. (2012) has suggested the integration of  
101 multiple climate parameters, with focus on reconstructing acute and widespread

102 synoptic deviations, rather than individual meteorological parameters for long tree-ring  
103 stable isotope records.

#### 104 1.2.2 Subfossil trees and stable isotope records

105 The availability of robust tree-ring chronologies is limited prior to the Holocene and  
106 only a few subfossil records have been established for the LG (e.g. Reinig et al. 2018  
107 and references therein). Absolutely dated tree-ring chronologies begin in 12 325 cal  
108 BP with the Preboreal Pine Chronology (PPC) from Germany followed by the  
109 Holocene Oak Chronology (HOC) (Friedrich et al. 2004). Beyond 12 325 cal BP,  
110 floating tree-ring chronologies exist for Switzerland (Reinig et al. 2018), Germany (e.g.  
111 Friedrich et al. 1999, 2001, 2004), France (Miramont et al. 2000a, 2000b, 2011), Italy  
112 (Casadoro et al. 1976), USA (Leavitt et al. 2006, Panyushkina and Leavitt 2007, 2013),  
113 New Zealand (Hogg et al. 2013, 2016) and Tasmania (Barbetti et al. 2004, Hua et al.  
114 2009). Scots pine trees from a floating tree-ring chronology (Reinig et al. 2018) was  
115 used for this study.

116 Some stable isotope records exist from subfossil tree-ring sequences from the  
117 prehistoric Holocene (Aguilera et al. 2011, Boettger, et al. 2003, Edvardsson, et al.  
118 2014, Frumkin 2009, Helama, et al. 2015, 2018, Helle 1994, Helle and Schleser 1998,  
119 Hunter, et al. 2006, Leavitt, et al. 2006, Mayr et al. 2003) and very few from the LG  
120 (Pauly et al. 2018, Wagner 2010, Becker et al. 1991). Many of these records lacked  
121 the sample replication to build a chronology (Aguilera et al. 2011, Frumkin 2009,  
122 Hunter et al. 2009), or developed a chronology with a limited temporal resolution or  
123 scale (Becker et al. 1991, Evardsson et al. 2013, Helama et al. 2015, Leavitt et al.  
124 2006, Mayr et al. 2003,). Only a few records exist with robust chronologies at annual  
125 (Pauly et al. 2018, Wagner et al. 2010) or decadal (Helama et al. 2018) resolution

126 during this time. Higher sample replication ( $n_{\text{trees}} > 5$ ) has been possible at decadal  
127 resolution compared to lower sample replication ( $n_{\text{trees}} = 1-4$ ) at annual resolution due  
128 to the limited availability and quality of subfossil wood material (See Supplementary  
129 Information 1).

## 130 **2. Material and Methods**

### 131 2.1 Stable Isotopes from pine trees in Switzerland

132 To develop a new tree-ring stable isotope chronology (hereafter referred to as CH-  
133 ISO) wood material from the improved Swiss Late Glacial Master Chronology, SWILM  
134 (Kaiser et al. 2012, Schaub et al. 2008, Friedrich et al. 1999) was used. While still  
135 floating, SWILM is the most robust and oldest LG tree-ring chronology currently  
136 available in the Northern Hemisphere, representing a very well-preserved LG forest. It  
137 covers 1500 years (~14 200 – 12 635 cal BP, Reinig et al. 2018) and comprises of  
138 trees from four pine stands in the Zürich region: Gaenziloo ( $n_{\text{trees}}=55$ ), Landikon  
139 ( $n_{\text{trees}}=28$ ), Daetttau ( $n_{\text{trees}}=46$ ) and Binz ( $n_{\text{trees}}=187$ ). The tree-ring chronology is  
140 plotted according to a local timescale “Zürich Scale” (representing floating positions  
141 prior to radiocarbon dating) as well as under cal BP. In conjunction with several  
142 hundred new  $^{14}\text{C}$  measurements of high quality - not only in terms of replication, but  
143 also resolution (annual) (Reinig et al. 2020) - the relative positioning of the SWILM  
144 with respect to IntCal13 (Reimer et al. 2013) was greatly improved (Reinig et al. 2018),  
145 resulting in a dating uncertainty of  $\pm 8$  years ( $2\sigma$  error). By independent inter-  
146 hemispheric  $^{14}\text{C}$  wiggle matching with New Zealand’s subfossil Kauri (Hogg et al.  
147 2016), the consolidated SWILM used in this study was shifted by +35 years with  
148 respect to IntCal13 (Sookdeo et al. 2019).

149 A subset of 27 trees was chosen from the SWILM wood material to establish a pair of  
150 oxygen and carbon stable isotope chronologies from tree-ring cellulose at annual  
151 resolution (CH-ISO, Figure 1), covering approximately 14 050 – 12 795 cal BP.  
152 Cellulose was extracted from wholewood material of individual tree-rings (Wieloch et  
153 al. 2011, Schollaen et al. 2017), homogenised and freeze-dried prior to being weighed  
154 and packed (silver capsules ( $\varnothing$ 3.3x4mm) for stable isotope measurement (Delta V,  
155 Thermofisher Scientific Bremen; coupled with TC/EA HT at 1400°C). Results were  
156 compared against international and lab-internal reference material (IAEA-CH3, IAEA-  
157 CH6 and Sigma-Aldrich Alpha-Cellulose) using two reference standards with  
158 widespread isotopic compositions for a single-point normalisation (Paul, Skrzypek,  
159 and Forizs 2007). Final isotope ratios are given in  $\delta$  value, relative to VSMOW ( $\delta^{18}\text{O}$ )  
160 and VPDB ( $\delta^{13}\text{C}$ ), with replication reproducibility of  $\pm 0.3\text{‰}$  ( $\delta^{18}\text{O}$ ) and  $\pm 0.15\text{‰}$  ( $\delta^{13}\text{C}$ ).

## 161 2.2 Data Analysis

162 Tree-ring records ( $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$ ) were z-scored, as outlined in *method one* from  
163 Hangartner et al. (2012) due to the high variability in absolute levels (‰) of individual  
164 trees. Pearson's correlations between (1) intra-tree isotope series of  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$   
165 (C:O) and (2) inter-tree  $\delta^{18}\text{O}$  were completed to investigate whether carbon and  
166 oxygen varied in tandem (thus controlled by similar environmental conditions) and to  
167 what degree  $\delta^{18}\text{O}$  was consistent across concurrently growing trees (i.e. the  
168 population signal). Inter-tree  $\delta^{18}\text{O}$  correlations were compared with sample replication  
169 using a linear model to explore the potential chronological bias.

170

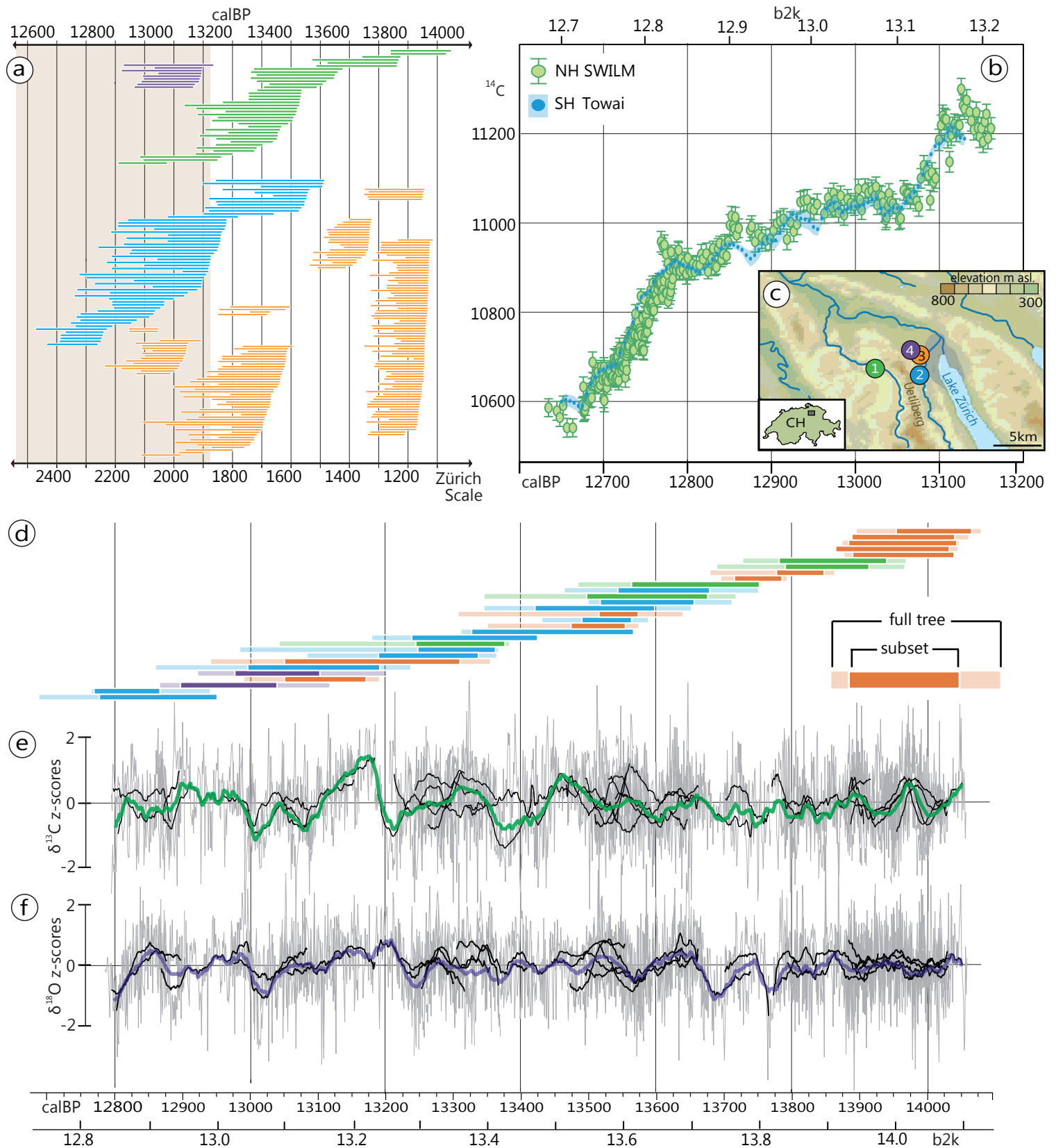


Figure 1. Subfossil trees in Switzerland: (a) Temporal distribution of a subset of subfossil tree-ring chronologies from 4 Swiss sites that make up the SWILM chronology: (1) Landikon (green,  $n_{\text{trees}} = 28$ ), (2) Gaenziloo (blue,  $n_{\text{trees}} = 55$ ), (3) Binz (orange,  $n_{\text{trees}} = 170$ ) and (4) Daetttau ( $n_{\text{trees}} = 9$ ). Additional groups of Binz and Daetttau unrelated to CH-ISO are not pictured. Site locations in (c) map, coloured according to (a). Trees were dated using high resolution radiocarbon (NH SWILM), wiggle-matched with decadal Southern Hemisphere Towai chronology. (b) Displays a subset of highest resolution dating results (timespan highlighted in a). (d) Tree replication for stable isotope chronology (CH-ISO). The 'full tree' represents the entire length of sample where tree-ring widths have been measured and cross-dated (Reinig et al. 2018, Kasier et al., 2012) and 'subset' indicates rings from full tree used for stable isotope analysis. Resulting individual stable isotope chronologies of carbon (e) and oxygen (f) with individual annual resolution in grey, 20-year individual means in black and chronology mean in green (carbon) and purple (oxygen).



171 **3. Results**172 3.1 Tree-ring stable isotopes

173 The isotope chronologies are a bit shorter than the SWILM (Figure 1d) due to  
174 limitations in cutting tracks and rings from the delicate subfossil wood material and  
175 avoidance of tree piths (to reduce any juvenile effects), with a total extent of ~14 050  
176 – 12 795 cal BP. The trees were chosen based on various selection criteria, including  
177 clearly expressed wood structure (allowing dissection of rings; ring-width  $\geq 0.3\text{mm}$ ), a  
178 relatively long lifespan ( $>150$  years), suitability for tree-ring cross-dating, and detailed  
179 radiocarbon series data. Despite this, many of the tree disks exhibited signs of  
180 degradation and decay, which became apparent during the dissection process;  
181 particularly along the outer rings comprising the sapwood which is softer than the inner  
182 heartwood and where the tree stumps were likely exposed to aqueous conditions.  
183 Approximate level of decay was noted during the sample preparation (Table 1),  
184 however, sections of very clearly visible decay and crumbled wood were avoided  
185 where possible during sample collection. The wood texture varied greatly between  
186 trees of all decay states – from brittle and dry to hard and resinous - but this was not  
187 quantitatively measured for this study. The sample replication varies throughout CH-  
188 ISO, with a maximum of 7, a minimum of 1 and a mean of 4.

189 3.2 Time Series Analysis

190 The inter-tree correlation of  $\delta^{18}\text{O}_{\text{tree}}$  was highly variable (Figure 2a), with a mean of  
191 approximately 0.18, with average values ranging from 0.01 to 0.41. A significant  
192 relationship was found between the average inter-tree  $\delta^{18}\text{O}$  correlation (Figure 2e) and  
193 the sample replication ( $r$ , Figure 2f) with an error of  $\pm 0.01$ :

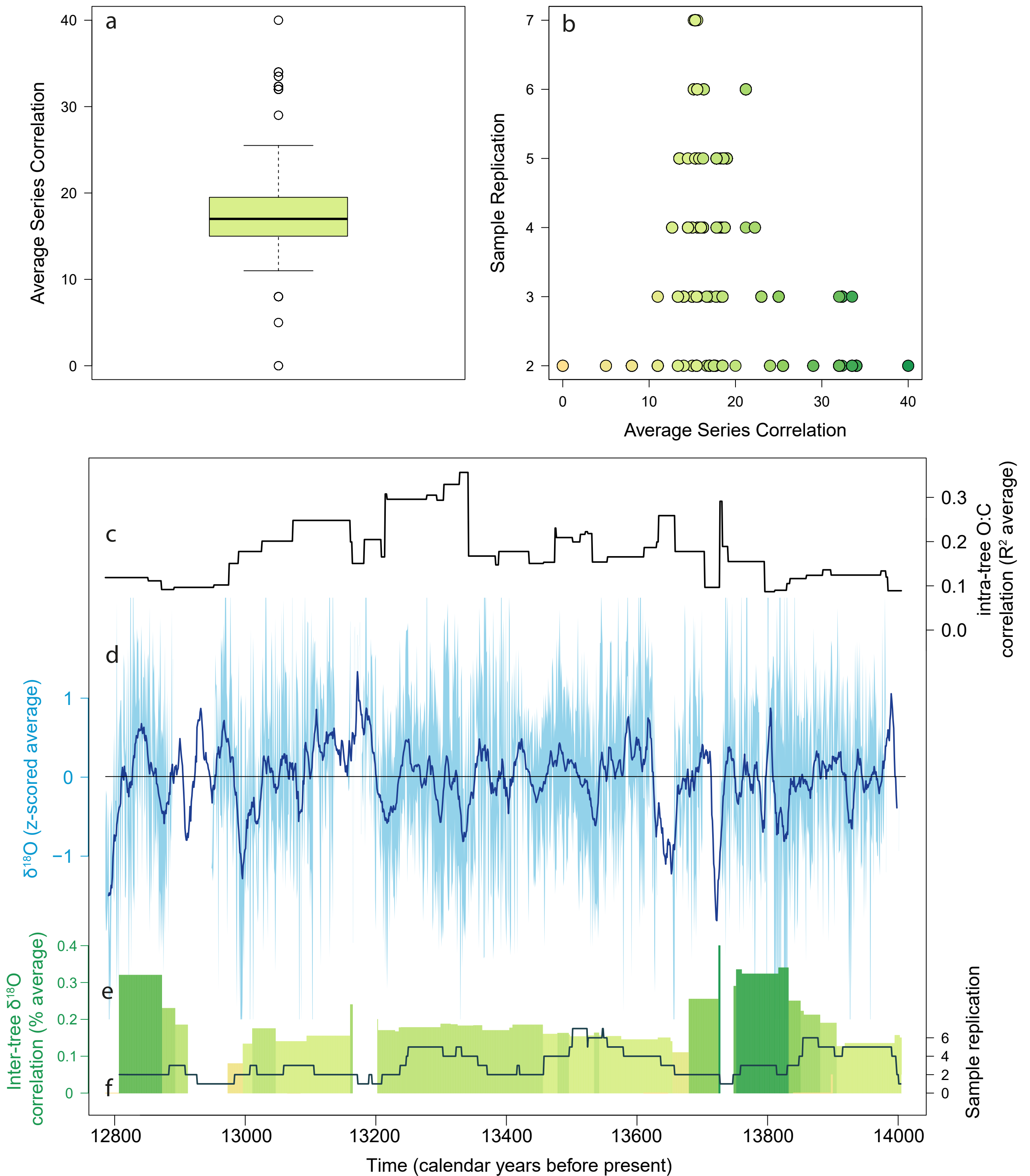


Figure 2. CH-ISO Chronology parameters: average series correlation of  $\delta^{18}\text{O}$  (a) boxplot and (b) compared to sample replication; (c) average individual (intra-tree) correlations between stable isotopes ( $\delta^{18}\text{O}:\delta^{13}\text{C}$ ), (d) mean tree-ring  $\delta^{18}\text{O}$  (z-scored) with standard deviation shaded, (e) average chronology (inter-tree)  $\delta^{18}\text{O}$  correlation, shaded from high correlation (green) to low correlation (yellow), and (f) chronology sample replication.

194 CH-ISO  $\delta^{18}\text{O}$  correlation =  $13.9 + 0.92*r \pm \text{error}$

195 This relationship demonstrates that sample replication is positively related to inter-tree  
196  $\delta^{18}\text{O}$  corrections. However, this linear relationship may be biased by more data points  
197 and spread in lower replication values (e.g. rep = 2 & 3, Figure 2b). For much of the  
198 chronology, the inter-tree  $\delta^{18}\text{O}$  corrections average between 12-20% (Figure 2b,e).  
199 The highest average inter-tree  $\delta^{18}\text{O}$  correlations (>25) occur in two distinct periods  
200 (~13 820 – 13 690 cal BP and 12 870 – 12 800 cal BP) with sample replications  
201 between 2-3 trees; demonstrating instances when outliers contradict the  
202 abovementioned linear relationship.

203 On the other hand, inter-tree correlation of  $\delta^{18}\text{C}_{\text{tree}}$  were generally very low (<0.10),  
204 and therefore further analysis will focus more on the suitability of  $\delta^{18}\text{O}_{\text{tree}}$  as a  
205 paleoproxy.

206 The covariance between the mean  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  is insignificant for the majority of  
207 trees within the chronology (Figure 2c). Only 9 of the 27 trees (33%) in CH-ISO  
208 demonstrated a  $\delta^{18}\text{O}:\delta^{13}\text{C}$  correlation  $R^2$  value >0.10.

209 15 trees are strongly decayed (Table 1 “high”: 56% CH-ISO), 9 trees show moderate  
210 decay (Table 1 “medium”: 33% CH-ISO) and 3 trees have little decay (Table 1 “low”:  
211 11% CH-ISO). Of those trees with significant inter-isotope correlations ( $R^2 >0.10$ ), one  
212 had “high” decay, 5 had “medium” decay and 3 had “low” decay. Therefore, 100% of  
213 the trees that exhibited low levels of decay, 56% with moderate decay and 7% with  
214 high decay showed strong inter-isotope correlations. Of these trees, the highest  
215 correlations were found in low decay trees (mean  $R^2 = 0.39$ ,  $n=3$ ) compared to  
216 moderately decayed (mean  $R^2 = 0.16$ ,  $n=5$ ) and highly (mean  $R^2 = 0.15$ ;  $n=1$ ) decayed  
217 trees. In addition, the vast majority of the highly decayed trees (93%) had low inter-

218 isotope correlations (mean  $R^2 = 0.03$ ,  $n=14$ ). When comparing individual tree  $\delta^{18}\text{O}$  to  
 219 overlapping trees in the chronology, there was no significant difference in the average  
 220 cross-correlation of different decay states – with a mean of 17.6% for low decay, 18.2%  
 221 for moderate decay and 19.6% for high decay. The cross-correlations varied  
 222 significantly, between 2% to 75% between tree pairs, averaging 7% to 40% for the  
 223 population compared with each individual (Table 1, Figure 2e).

Sample	$R^2$ (O:C)	P (O:C)	Decay	Average inter-tree cross correlation (%)	Timespan (cal BP)	Zurich Scale
LAN10	0.15	<0.001	High	18	13355 - 13241	1740-1854
GAE28	0.06	<0.05	High	20	13330 - 13183	1765-1912
GAE67	0.16	<0.001	Medium	14	13303 - 13045	1792-2050
K353	0.15	<0.001	Medium	24	13185 - 12991	1910-2104
K352	0.15	<0.001	Medium	7	13165 - 13005	1930-2090
GAE59	0.02	-	High	11	13091 - 12967	2004-2128
GAE60	0.01	-	High	5	12984 - 12842	2111-2253
GAE25	0.06	<0.05	High	32	12883 - 12795	2212-2309
GAE5	0.001	-	High	32	12904 - 12795	2192- 2365
LAN44	0.001	-	High	30	13904 - 13781	1191-1314
BIN138	0.19	<0.001	Medium	29	13838 - 13768	1257-1327
BIN137	0.01	-	High	40	13767 - 13697	1328-1398

LAN21	0.16	<0.001	Medium	11	13743 - 13555	1352-1540
GAE68	0.05	<0.05	Medium	17	13670 - 13535	1425-1560
LAN16	0.07	<0.001	Medium	17	13667 - 13489	1428-1606
GAE74	0.015	-	High	13	13647 - 13 509	1448-1586
GAE48	0.02	0.1	High	19	13589 - 13413	1506-1682
BIN150	0.42	<0.001	Low	14	13565 - 13 507	1530-1588
BIN75	0.14	<0.001	Low	18	13558 - 13320	1537-1775
GAE45	0.01	-	High	11	13546 - 13466	1549-1629
GAE51	0.001	-	High	16	13417 - 13232	1678-1863
GAE31	0.6	<0.001	Low	21	13369 - 13238	1726-1857
BIN155	0.01	-	Medium	7	14022 - 13855	1073-1240
BIN162	0.06	<0.001	High	14	14033 - 13874	1062-1221
BIN191	0.07	<0.001	High	17	14030 - 13879	1065-1216
BIN207	0.003	-	High	16	14050 - 13944	1040-1151
LAN37	0.001	-	Medium	38	13930 - 13772	1165-1323

224 Table 1: Individual tree correlations between isotopes ( $\delta^{18}\text{O}:\delta^{13}\text{C}$ ) within a single tree,  
 225 average inter-tree  $\delta^{18}\text{O}$  cross-correlations and preliminary qualitative decay assessment of  
 226 each tree. Significant inter-isotope correlations are highlighted.

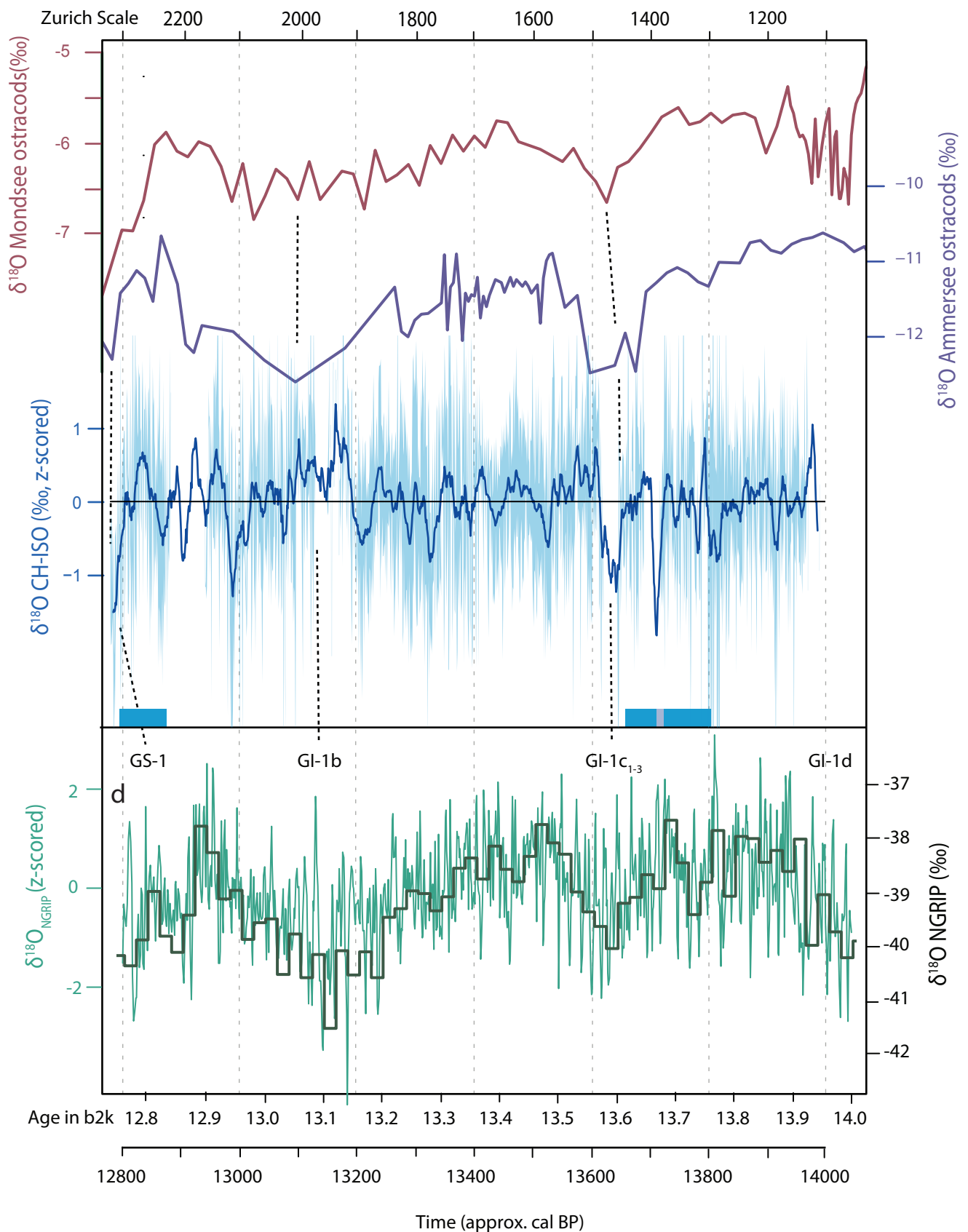


Figure 3. An overview of Late Glacial climate oscillations in the Northern Hemisphere from (A) Austria (Lake Mondsee, Lauterbach et al. 2011), (B) Germany (Lake Ammersee, von Grafenstein et al. 1999) and (D) Greenland (Rasmussen et al. 2014), compared to the Swiss tree-ring stable oxygen isotope record (CH-ISO, this study). Four NGRIP cool substages indicated (GI-1d, GI-1c<sub>1,2</sub>, GI-1b and GS-1) matched with comparable climate oscillations in other archives. Blue bars represent periods of relatively high population signal within the CH-ISO stable oxygen series.

## 227 **4. Discussion**

### 228 4.1 Considerations in CH-ISO chronology development

#### 229 4.1.1. Cellulose decay and stable isotopes

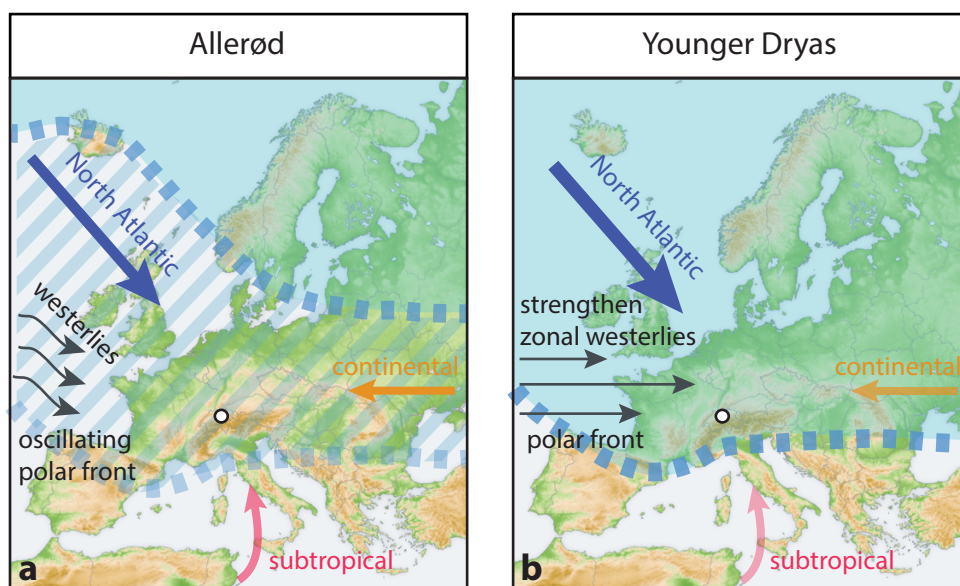
230 Due to wood and cellulose decay considerations (Supplementary Material 1), in  
231 addition to the fact that cellulose is non-labile and does not exchange oxygen isotopes  
232 with xylem water following formation (Wright et al. 2008), cellulose was extracted for  
233 the development of CH-ISO. Similar to Nagavciuc et al. (2018), there were  
234 divergences in both  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  between trees, with weaker population signals in  
235  $\delta^{13}\text{C}$  despite the use of standard sample replication. As a result of the low correlation  
236 of  $\delta^{13}\text{C}_{\text{tree}}$  between trees, this discussion will focus on the implications of  $\delta^{18}\text{O}_{\text{tree}}$ .

237 Evidence of decay can be seen around bark and sapwood in most CH-ISO trees  
238 (Figure 4), where the wood material is relatively discoloured and brittle, indicating  
239 variable levels of rot. Based on other studies (Savard et al. 2012; Blanchette 2000),  
240 we assume this is a result of bacteria from water-saturated soil, preferentially attacking  
241 the outer trunk portions through water infiltration. Yet, due to the expectation that  
242 stable oxygen in cellulose is less vulnerable to diagenetic alteration (Leavitt et al.  
243 1993), the varying decay states were not considered when developing the CH-ISO  
244 chronology. In CH-ISO, the subfossil trees exhibit highly variable states of visual decay  
245 (Table 1), as demonstrated by a qualitative assessment. While preliminary, it appears  
246 that trees with less extensive decay have significantly higher inter-isotope ( $\delta^{18}\text{O}$ :  $\delta^{13}\text{C}$ )  
247 correlations than those with moderate to high levels of decay. In cases where the  
248 stable isotopes do not exhibit similar trends, it is assumed that climatological variables  
249 which impact both isotopes through vapor pressure deficit and leaf stomatal  
250 conductance (e.g. relative humidity, precipitation) are not strongly recorded within the  
251 tree-rings. As a result, it can be assumed that trees with relatively high levels of decay





**Figure 4. Swiss subfossil pine wood quality** (a) disk of tree demonstrating variable levels of decay: (1) low, (2) moderate and (3) high. Higher levels of wood decay were generally found closer to the bark, with best preserved wood material closer to the pith (b).



**Figure 5. Hypothesised sequence of Late Glacial air mass teleconnections and migrations:** (a) general Allerød conditions with oscillating polar front; the Allerød representing the Interstadial in the Late Glacial between 13 900 - 12 900 cal BP, followed by (b) the Younger Dryas (GS-1) cold reversal, with increased zonal westerlies and migrated polar front. Study site indicated (white circle).



252 lead to more extensive chemical alteration of  $\delta^{13}\text{C}_{\text{tree}}$  over  $\delta^{18}\text{O}_{\text{tree}}$ , causing the greater  
253 divergence between the two measures and between  $\delta^{13}\text{C}_{\text{tree}}$  of different trees.

254 Inter-tree  $\delta^{18}\text{O}_{\text{tree}}$  correlations did not significantly differ as a function of wood decay,  
255 further suggesting that  $\delta^{18}\text{O}_{\text{tree}}$  may be less prone to fractionation during wood decay,  
256 or at least wood decay that is clearly visible. The still relatively low cross-correlation  
257 between inter-tree  $\delta^{18}\text{O}_{\text{tree}}$  (2% - 75%; mean: 15%) implies that chemical wood decay  
258 may be present but indistinguishable under light microscope. This is within the range,  
259 if not slightly lower than the inter-tree  $\delta^{18}\text{O}_{\text{tree}}$  correlations for modern pine trees in the  
260 region; a cross-correlation of 29% ( $n_{\text{trees}} = 4$ ) was found by Reynolds-Henne et al.  
261 (2007) in the Swiss alps.

262 In accordance to our findings, Savard et al. (2012) established divergences in  $\delta^{18}\text{O}$   
263 between trees in the absence of visual decay, proving further processes of aqueous  
264 decay impacting subfossil wood are yet to be discovered. Indeed, while strongly  
265 decayed portions of tree disks were avoided where possible for stable isotope  
266 analysis, some differences in the structural integrity of the wood (in the absence of  
267 colouration change) was noted. The impact of textural variability on tree-ring isotopes  
268 was not taken into consideration and requires further investigation.

#### 269 4.1.2 Local site and tree-level considerations

270 Local site conditions, which may impact tree-ring isotopic signatures, are difficult to  
271 decipher for CH-ISO. Trees from SWILM grew along an unstable slope, impacting root  
272 development and potentially uptake of sourcewater. Furthermore, varying depths of  
273 the active soil layer from (potential) permafrost conditions during cold episodes could  
274 have caused distinct variability in subsurface hydrology and rooting depth. Fluctuating  
275 thaw depth could have provided inconstant amounts of melted permafrost water to

276 individual trees, and it can well be assumed that the soil water  $\delta^{18}\text{O}$  with a large  
277 contribution from permafrost melt had a much different  $\delta^{18}\text{O}$  signature (more depleted)  
278 than summer rain. Moreover, the presence or absence of permafrost could have  
279 influenced the seasonal uptake of different water sources with various isotopic  
280 signatures (e.g. relatively  $\delta^{18}\text{O}$ -enriched liquid rain vs.  $\delta^{18}\text{O}$ -depleted snow). For  
281 example, meltwater and/or permafrost thaw could lead to depleted  $\delta^{18}\text{O}$  signatures  
282 despite high summer temperatures. Potentially drier and warmer summer conditions  
283 may lower soil water  $\delta^{18}\text{O}$  values, i.e. altering tree water source composition due to  
284 increased permafrost or snowpack meltwater. This can lead to inverse climate- $\delta^{18}\text{O}_{\text{cel}}$   
285 relationships as indeed reported for larch trees in Siberia (Saurer et al 2016).  
286 Furthermore, during very dry summers meltwater may have served as the most  
287 important water source for the trees, potentially compensating increased  $^{18}\text{O}$   
288 enrichment of needle water due to increased transpiration rates during drought years.

289 As a result of these potential differences in microsite-conditions, trees within the same  
290 vicinity could theoretically have divergence sourcewater signals. For example, more  
291 stable soil may have longer precipitation water residence times and trees may be less  
292 prone to take up water from event precipitation or subsequent surface runoff (which  
293 have different isotopic values) (Genereux & Hooper 1998). It is therefore expected that  
294 trees growing in unstable subsites may track short-term hydrological events more  
295 effectively (higher inter-annual  $\delta^{18}\text{O}$ ) than those in more stable subsites. At present, it  
296 is impossible to tell which trees diverge from a population signal due to slope instability  
297 specifically without significantly increasing the sample replication.

298 Scenario-specific responses of tree-rings to meteorological drivers may be an  
299 important factor moderating the stability between subfossil tree-ring  $\delta^{18}\text{O}$  and regional

300  $\delta^{18}\text{O}$  correlations at high frequency over the LG. For example, during dry periods, trees  
301 tend to reduce stomata conductance to limit increasing water loss by transpiration due  
302 to increased leaf-to-air vapour pressure difference. Dry summers usually lead to  
303 increased  $^{18}\text{O}$  enrichment of leaf water and correspondingly higher  $\delta^{18}\text{O}_{\text{tree}}$  values  
304 compared to sourcewater values (e.g. Kahmen et al. 2011). Conversely, during humid  
305 periods,  $^{18}\text{O}$  enrichment is reduced leading to reduced  $\delta^{18}\text{O}_{\text{tree}}$  values being closer to  
306  $\delta^{18}\text{O}$  of source (assuming oxygen isotope exchange between xylem water and sugars  
307 during cellulose biosynthesis in the trunk is constant during dry and wet years).  
308 Therefore, in times of high precipitation, regional  $\delta^{18}\text{O}$  precipitation tends to be  
309 incorporated more readily into tree-ring cellulose, with less overprinting due to  
310 physiologically driven fractionation, which may be the case during the periods of strong  
311  $\delta^{18}\text{O}$  population signals and associated  $\delta^{18}\text{O}$  depletions within CH-ISO.

#### 312 4.1.3 Temporal stability of CH-ISO

313 Time series analysis revealed high variability in the strength of the  $\delta^{18}\text{O}$  population  
314 signal, with a linear model suggesting higher sample replication leads to higher inter-  
315 tree  $\delta^{18}\text{O}_{\text{tree}}$  correlations (Results 3.2). However, numerous outliers (Figure 2a)  
316 suggest this relationship is limited. For example, relatively high correlations (>25%)  
317 exist only during periods of 2-3 tree replication (Figure 2b) and strong  $\delta^{18}\text{O}_{\text{tree}}$   
318 depletions (Figure 2d), despite moderate wood decay. During these extremes, intra-  
319 tree  $\delta^{18}\text{O}_{\text{tree}}:\delta^{13}\text{C}_{\text{tree}}$  correlations are generally quite low (Figure 2c,  $R^2 < 0.20$ ), with  
320 exception to the  $\delta^{18}\text{O}_{\text{tree}}$  extreme at ~13 765 cal BP. This reveals that when the CH-  
321 ISO trees exhibit significantly similar  $\delta^{18}\text{O}_{\text{tree}}$  values, the  $\delta^{18}\text{O}_{\text{tree}}$  is generally not well  
322 correlated to  $\delta^{13}\text{C}_{\text{tree}}$ . Therefore, during these periods, the environmental factors that  
323 have the potential to modulate both isotopes (e.g. relative humidity, temperature) are

324 not strongly recorded in the tree-rings; but rather, factors that only control  $\delta^{18}\text{O}_{\text{tree}}$   
325 (precipitation source, type and amount) are possibly dominating the population signal  
326 and/or wood decay is preferentially altering the  $\delta^{13}\text{C}_{\text{tree}}$ . Note, temporal instability of  
327 climate-isotope relationships are also observed in modern stable isotope records from  
328 Switzerland (Reynolds-Henne et al. 2007). Potentially, uncertainties in climate-proxy  
329 relationships can be detected by comparing independent proxy records from different  
330 archives, provided that they generally reflect the same climate variable at similar time  
331 and spatial resolution.

#### 332 4.2. The potential of subfossil tree-ring $\delta^{18}\text{O}$ to record hydroclimate variability

333 Swiss Plateau sourcewater ( $\delta^{18}\text{O}$ ) is broadly driven by precipitation originating from  
334 North Atlantic air masses but can also receive other air masses (sub-tropical,  
335 continental; Figure 5, Kozel and Schotterer, 2003) and local hydrological variations  
336 influenced by geomorphological (local hillslope) processes. This demonstrates an  
337 element of non-stationarity to the tree-ring isotope relationship to climate variables  
338 over time, which has also been found in modern tree-ring isotope records (Kress et al.  
339 2010, Reynolds-Henne et al. 2007, Seftigen et al. 2011). Shorter-term fluctuations in  
340  $\delta^{18}\text{O}$  precipitation correlate to synoptic patterns in precipitation-bearing air masses  
341 over temperature (Gat & Carmi 1987), overriding the general temperature-precipitation  
342 relationships and are further affected by leaf- and soil- level evaporative enrichment  
343 during warm/dry periods. As a result, rapid changes in precipitation isotopes may be  
344 strong, yet independent of temperature, creating significant deviations from the classic  
345 precipitation-temperature Dansgaard relationship (Edwards et al. 1996, Teranes and  
346 McKenzie 2001, Hammarlund et al. 2002).

347 Swiss water isotopes recorded in tree-rings are theoretically sensitive to summer  
348 (growing season; JJA) precipitation, whereas Greenland water isotopes ( $\delta^{18}\text{O}_{\text{NGRIP}}$ )  
349 are more strongly driven by winter precipitation and temperature (DJF). Similarly,  
350 water isotopes from LG southern France speleothems reflect winter (recharge season)  
351 precipitation; the signal of which is also moderated by summer evaporation and  
352 deviations in storm track pathway (e.g. Genty et al. 2006). Furthermore, regional lake  
353 water isotopes are sensitive to annual temperature variations and subject to reservoir  
354 effects (e.g. von Grafenstein 2000). These isotopic differences demonstrate the  
355 inherent seasonality of these archives, which makes regional climate change  
356 comparison challenging. Moreover, circulation controls the  $\delta^{18}\text{O}$  signal in the  
357 atmosphere more strongly during the winter (due to stronger temperature gradients  
358 and less extreme continental moisture cycling; Baldini et al. 2008), making winter  $\delta^{18}\text{O}$   
359 a more sensitive recorder of changes in atmospheric circulation. Accordingly, annual  
360 or winter driven proxy records (e.g. lakes, ice cores, speleothems) may have more  
361 success reconstructing such hemispheric variability compared to summer proxies (e.g.  
362 trees), unless the circulation change is extreme and/or prevails for a few consecutive  
363 years (Saurer et al. 2012). Modern  $\delta^{18}\text{O}_{\text{tree}}$  reveals the strongest climate correlations  
364 during circulation changes (Saurer et al. 2012), suggesting that widespread climate  
365 extremes could be reconstructed from LG  $\delta^{18}\text{O}_{\text{tree}}$ .

#### 366 4.3 Challenges in reconstructing LG climate oscillations from the CH-ISO subfossil 367 tree-rings

368 Due to complex and dynamic factors influencing the propagation of  $\delta^{18}\text{O}$  from the  
369 atmosphere into tree-ring cellulose, long-term and widespread climate interpretation  
370 from subfossil dendroisotope records is challenging; particularly when sample

371 selection is limited. For example, while two drops in  $\delta^{18}\text{O}_{\text{tree}}$  seem to approximately  
372 parallel NGRIP 'cool periods' ( $\delta^{18}\text{O}_{\text{NGRIP}}$  depletions; Figure 3), the inter-annual  
373 variability as well as low frequency trends of the rest of the dataset show limited visual  
374 resemblance between the records. The fluctuating similarity between tree-ring  $\delta^{18}\text{O}$   
375 chronologies (e.g. Pauly et al. 2018 and this study) and other regional LG  $\delta^{18}\text{O}$  records  
376 prove climate-proxy relationships in subfossil dendroisotope records are not stable  
377 through time – shifting between recording regional and local weather conditions as  
378 well as being influenced by tree- and population- level factors.

#### 379 4.4 Low frequency $\delta^{18}\text{O}$ similarity between LG archives

380 Negative excursions ('cool periods') present in both European and Greenland  $\delta^{18}\text{O}$   
381 records (GI-1C, GS-1c) are evident in  $\delta^{18}\text{O}_{\text{tree}}$  (Figure 3) as significant depletions.  
382 These occur when inter-tree  $\delta^{18}\text{O}_{\text{tree}}$  correlations are high, and thus the  $\delta^{18}\text{O}_{\text{tree}}$   
383 chronology is more likely to be recording regional climate conditions over local effects.

384 This connection between  $\delta^{18}\text{O}_{\text{tree}}$  and other records (Lauterbach et al. 2011,  
385 Rasmussen et al. 2014, von Grafenstein et al. 1999) could be due to an increased  
386 prevalence of North Atlantic air masses in mainland Europe as a result of a southerly  
387 moving polar front (e.g. Kageyama et al. 1999, Schenk et al. 2018). Models have  
388 predicted a winter amplification of this process due to sea ice expansion in winter and  
389 atmospheric blocking of westerlies in summer during the Late Glacial (e.g. Schenk et  
390 al. 2018 and references therein), which may explain its muted signature in the summer  
391 tree-ring archive for CH-ISO in comparison to the lake and ice core records.

392 On the other hand, one other Greenland isotope substage (GI-1b) - demonstrated as  
393 the Gerzensee Oscillation across Europe (e.g. Eicher & Siegenthaler 1976,  
394 Lauterbach et al. 2011, Lotter et al. 1992, 2012) - is not identifiable in CH-ISO. This

395 could be due to the relatively low tree correlation (5-15%) in  $\delta^{18}\text{O}_{\text{tree}}$  over this interval  
396 as a result of local site conditions overriding regional climate signals.

397  $\delta^{18}\text{O}$  records display a distinct excursion beginning between ~12 900 – 12 600 cal BP  
398 (Figure 3), ascribed as the onset of the GS-1 / Younger Dryas. A short tree-ring  
399 sourcewater reconstruction from southern France (Barbiers: ~12 900 – 12 600 cal BP,  
400 Pauly et al. 2018) revealed an onset at 12740 cal BP, reflected by an increasing  
401 influence of both Mediterranean (+  $\delta^{18}\text{O}_{\text{tree}}$ ) and North Atlantic (-  $\delta^{18}\text{O}_{\text{tree}}$ ) storms,  
402 attributed to an oscillating polar front and more extreme weather locally. In contrast,  
403 SWILM trees exhibit high inter-annual variability along a low frequency decline (-5.5‰  
404 over 40 years), providing further evidence of unstable weather conditions during the  
405 cooling onset (Steffenson et al. 2008). Subfossil forests from both Barbiers (Pauly et  
406 al. 2018) and SWILM (Reinig et al. 2018; Figure 2a) demonstrate a distinct decline in  
407 excavated trees at approximately ~12 600 cal BP, suggesting a tree die off concurrent  
408 with the sustained drop in all  $\delta^{18}\text{O}$  records.

## 409 5. Conclusions

410 The subfossil tree-ring  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  records developed in this study reflect a complex  
411 set of interacting climate variables modulated by local factors (microclimate, wood  
412 degradation); all of which must be carefully identified and investigated prior to a  
413 comprehensive paleoclimate interpretation.

414 Detailed research in the calibration of cellulose decay and related isotopic fractionation  
415 is required as a prerequisite to further developing this chronology in order to increase  
416 the population signal of both  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$ . In particular, experiments on present-day  
417 wood samples infected with certain fungi and/or bacteria targeting the isotopic effects  
418 of cellulolytic enzymes are recommended. Furthermore, a suite of pre-treatment

419 protocols would have to be completed, including cellulose content weighing (to  
420 measure mass loss of whole wood), wood strength tests (to determine texture) as well  
421 as wood anatomy and colorimetry (to distinguish decay), prior to selecting optimal  
422 subfossil samples for further analysis. Also, additional isotope data from wholewood,  
423 cellulose and/or lignin methoxyl groups will potentially allow to better identify  
424 problematic sequences in multi-parameter approach (e.g. Mischel et al. 2015). As  
425 Swiss subfossil trees continue to be dated (Reinig et al., 2018), a revised subfossil  
426 dendroisotope protocol and continued expansion of CH-ISO will follow.

427 The strongly variable (visual) correlation between NGRIP, Mondsee, Ammersee and  
428 CH-ISO across millennia demonstrates that proxy-climate relationships are complex  
429 and non-stationary through time, beyond the differences in temporal resolution and  
430 dating uncertainties. Despite having the same variable ( $\delta^{18}\text{O}$ ), each dataset records  
431 very different aspects of environmental conditions across space and time. A multi-  
432 proxy approach with regional inter-archive comparisons is vital in detangling the  
433 relative contribution of different factors in the overall local isotopic signal. Nonetheless,  
434 this new stable isotope record here demonstrates the potential to hydroclimate  
435 conditions in the Late Glacial using subfossil pine trees.



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444 Competing Interests Statement

445 We declare no competing interests in this work

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### 1. Wood decay and tree-ring stable isotopes

Wood decay and decomposition have been shown to significantly impact intra- and inter- tree-ring stable isotopes, particularly in subfossil wood (e.g. Ziehmer et al. 2016). Diagenetic effects, in terms of degradation from biotic (bacteria, fungi) and abiotic (hydrological) decay, impact the structural and therefore the chemical nature of wood (Björddal et al. 1999, Daniel 1994, Benner et al. 1987, Daniel & Nilsson 1998, Staccioli et al. 1997, Schleser et al. 1999). It has been well established that cellulose is preferentially decayed over lignin and other wood components, impacting their relative ratios in wholewood over time (Park & Epstein 1961, Spiker & Hatcher 1987, Schleser 1999). This differential decay is a crucial consideration, as various wood components have divergent isotopic signatures, which can significantly impact the resultant wholewood stable isotopic signature. For example, a consistent enrichment of  $\delta^{13}\text{C}$  (Loader et al. 2003; Lukens et al. 2019) and inconsistent  $\delta^{18}\text{O}$  contained in lignin compared to cellulose (Park & Epstein 1961, Wilson & Grinsted 1977, Barbour 2002, McCarroll & Loader 2004, Borella et al. 1999). During burial and preservation of subfossil wood, infiltration of sediment-laden water leads to the reduction of specific acidic groups (native carboxyls and salt-derived carboxylates), and production of others (e.g. catalysation of esters) (Menchi et al. 1997, Staccioli et al. 1994). Hence, in this study cellulose was extracted in an attempt to avoid isotopic artefacts arising from the use of wholewood.

Apart from cellulose decay (due to burial post-death spread fungi and maybe more importantly bacteria), oxygen and carbon ions can be redistributed between tree-rings (English et al. 2011), significantly impacting inter-annual variability and particularly isotopic trends from pith to bark (and/or heartwood to sapwood). Decay has shown

instances of  $\delta^{13}\text{C}$  depletion proportional to diminishing cellulose content to a specific threshold (Schleser et al. 1999), significant decreases in  $\delta^{18}\text{O}$  (Savard et al. 2012), as well as generally unreliable stable isotope values (Savard et al. 2012; Nagavciuc et al. 2018; Savard et al. 2012) within the same tree and between trees at the same site.

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