Millennial-scale deglaciation across the European Alps at the transition between the Younger Dryas and the Early Holocene - evidence from a new cosmogenic nuclide chronology

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1 Millennial-scale deglaciation across the European Alps at the
2 transition between the Younger Dryas and the Early Holocene –
3 evidence from a new cosmogenic nuclide chronology

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10 Moreau, L. & ASTER Team: Millennial-scale deglaciation across the European Alps at the
11 transition between the Younger Dryas and the Early Holocene – evidence from a new
12 cosmogenic nuclide chronology. *Boreas*....

14 Reconstructing the spatial and temporal response of mountain glaciers to rapid climate change
15 in the past provides access to the effects of current climate change. Yet, the spatial and temporal
16 variability of past glacier fluctuations is not fully understood. In this study, we focus on the
17 timing of glacier fluctuations in the European Alps during the Younger Dryas/Early Holocene
18 (YD/EH) transition. In an effort to elucidate whether glacier fluctuations were synchronous
19 during this period, we present a new chronology of the Alpine Talèfre glacier, based on 14 new
20 ¹⁰Be ages of moraines and roches moutonnées. The retreat of Talèfre glacier was initiated
21 during the mid-YD (~12.4 ka), then it experienced a gradual retreat punctuated by at least 3
22 oscillations until ~11 ka before shrinking substantially within its Little Ice Age limits (13th -
23 19th century). Comparison of our findings with published glacier chronologies in the Alpine
24 region highlights broadly synchronous behaviour of glaciers across the Alps between 12 and
25 10 ka. The coeval glacier fluctuations at a regional scale suggest that common regional climate
26 conditions had a major impact on Alpine glacier variations during the YD/EH transition. The
27 similarity of glacier behaviour and independent temperature records in both the Alpine region
28 and the northern high latitudes suggests a teleconnection between these regions, but differences

in the amplitude of the mean annual temperature signals relative to summer temperature indicate pronounced changes in seasonality between the YD and the EH.

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Mountain glaciers are recognized as sensitive recorders of climate variability, and their fluctuations mainly depend on variations of both temperature during the melt season and snow precipitation (Oerlemans 2005; Roe *et al.* 2017). In the case of Alpine glaciers, their recent evolution is controlled by large-scale climate variations moderated by local specific factors (Rabatel *et al.* 2013; Vincent *et al.* 2017). In the light of current climate evolution and rapid retreat of glaciers worldwide, it is important to improve our understanding of the sensitivity of glaciers to rapid climate changes in the past, such as those during the Younger Dryas/Early Holocene (YD/EH) transition (between ~12 and 10 ka). Therefore, the impact of large-scale

climate variations versus local climate and site-specific factors has to be better assessed by constraining past glacial chronologies.

Moraines and roches moutonnées represent clear geomorphic markers of past glacier extents. Studying and dating them is a valuable approach for constraining former glacier behaviour, establishing glacier chronologies and investigating the related climate changes. Due to the lack of direct dating methods, the traditional chronology of glacier advances (“stadials”) in the Alps during the period of Lateglacial and EH deglaciation is mainly based on comparative studies of moraine stratigraphies and morphologies and on the variation of the equilibrium line altitude (ELA) (e.g. Maisch 1981). Temporal correlations of these stadials with palaeoclimatic periods were broadly confirmed by the pioneering cosmogenic nuclide studies at a few Alpine sites (see reviews in Ivy-Ochs *et al.* 2006 and Ivy-Ochs 2015). For instance, the Egesen stadial moraines are assigned to the Younger Dryas (from 12.9 to 11.7 ka; Rasmussen *et al.* 2006), and are associated with an ELA depression of 250–350 m relative to the Little Ice Age (LIA) (Ivy-Ochs 2015). The Kartell stadial moraines are assigned to the Preboreal cold period (11.4 – 11.3 ka; Kobashi *et al.* 2008) with an ELA depression of 120 m (Ivy-Ochs 2015). However, recent studies showed that correlating moraines from different Alpine valleys solely based on the above-mentioned comparative criteria is unreliable, highlighting the need for direct moraine dating with cosmogenic nuclides when comparing moraines from various sites (e.g. Reitner *et al.* 2016; Boxleitner *et al.* 2019).

Over the past few years, parameters for the calculations of cosmic ray exposure ages were refined (e.g. Balco *et al.* 2008, 2009; Young *et al.* 2013; Borchers *et al.* 2016; Balco 2017; Martin *et al.* 2017) and an increasing number of cosmogenic nuclide dating studies from the Alps have been published, providing evidence of multiple glacier fluctuations during the Lateglacial and the EH (e.g. Schindelwig *et al.* 2012; Moran *et al.* 2016a, b; Baroni *et al.* 2017). Attempts of identifying temporal or spatial trends in Alpine glacier behaviour during the

Lateglacial and EH have been made in earlier studies (Schimmelpfennig *et al.* 2014; Chenet *et al.* 2016; Braumann *et al.* 2020). Although age uncertainties are generally too large to infer clear trends, an abrupt glacier recession is not documented in the Alps at the very sharp transition between YD and Holocene that is observed in Greenland ice cores with a $\sim 10^{\circ}\text{C}$ warming in less than one hundred years.

Rather, previous studies suggested that EH glacier stillstands or advances in the Alps could be a response to abrupt cold spells observed in Greenland ice cores (e.g. Ivy-Ochs *et al.*, 2009; Schindelwig *et al.* 2012; Schimmelpfennig *et al.* 2012, 2014), such as the Preboreal Oscillation or a cold event notable at ~ 10.9 ka (Rasmussen *et al.* 2006; Vinther *et al.* 2006).

In the Western Alps, glacier chronologies from Lateglacial and EH have for a long time been very scarce, but recent robust records and additional data from the French Alps will now provide the opportunity to use the Alpine data set as a whole for a regional analysis (e.g. Cossart *et al.* 2012; Chenet *et al.* 2016; Hofmann 2018; Protin *et al.* 2019; Schimmelpfennig *et al.* 2019). The chronological investigation of these moraine chronologies could lead to a better understanding of the large-scale palaeoclimatic conditions during a period for which proxy and model data still show diverging temperature trends in both mid- and high latitudes: coupled ocean-atmosphere climate models (e.g. Buizert *et al.* 2014) failed to reproduce the abrupt warming at the end of the YD found in North Atlantic and Greenland ice cores (e.g. Rasmussen *et al.* 2006; Shakun *et al.* 2015), whereas continental temperature reconstructions based on biological proxies either fit with this drastic warming (chironomid records in Heiri *et al.* 2015) or indicate a much more gradual warming (e.g. pollen records in Marsicek *et al.* 2018 and glacier chronologies in Rainsley *et al.* 2018, and Young *et al.* 2020).

In this study, we provide a new glacier chronology in the French Alps by analyzing the Talèfre glacier patterns during the deglaciation of the “Jardin de Talèfre”, one of the highest moraine sites in the Alps (close to 3000 m a.s.l.), where a set of moraines is deposited above

roches moutonnées. We present new *in-situ* cosmogenic beryllium-10 (^{10}Be) ages of moraines and roches moutonnées for this area and compare these new data with published ^{10}Be glacial chronologies from elsewhere in the European Alps, selected for their quality (e.g. number of samples, dispersion of the ages) and re-calculated using uniform methods. Building on this regional compilation, we evaluate (i) the possibility of a regional synchronicity of glacier fluctuations during the multi-millennia period of the end of the YD and the beginning of the Holocene (hereafter YD/EH transition), (ii) the impact of large-scale climate variations on glacier patterns versus site-specific factors, (iii) the underlying climatic drivers in a regional scale and their potential link with the high-latitudes, and (iv) the climatic significance of the small fluctuation range of glacier extent between the YD and EH in the light of the significant and abrupt climate change recorded in independent proxy records.

Study site

Talèfre glacier (45°54'N, 6°59'E) is a cirque glacier located in the French part of the Mont-Blanc massif and was a major tributary of Mer de Glace glacier, the largest glacier in the Mont-Blanc massif (Fig. 1). To the north is located the Argentièrè glacier, the second largest glacier in the Mont-Blanc massif (Fig. 1). In 2008, Talèfre glacier covered an area of ~8 km² with a maximum length of 4.5 km and altitude spanning 2473 to 3743 m a.s.l. (Gardent 2014). The ELA of Talèfre glacier for the period 1984-2010 was estimated to ~3000 m a.s.l. (Rabatel *et al.* 2013). Next to Argentièrè glacier, the mean annual temperature and precipitation for the time period 1979-2002 are estimated at 1.8 °C and 1783 mm at ~2000 m a.s.l. (Joly *et al.* 2018; D. Joly pers. comm. 2018). The catchment of Talèfre glacier is bordered by granite rocks of late Hercynian age (~300 Ma; Bussy *et al.* 2000).

During the YD, Talèfre glacier was connected to Mer de Glace, which at that time occupied the Arve Valley with a front located westward of the left boundary of Fig. 1 (Prud'homme *et al.* 2020). During the last LIA advance in the 19th century, Talèfre glacier was confluent with Leschaux glacier, which further downvalley joins Tacul glacier to form Mer de Glace (Fig. 1). Since ~1950 the Talèfre and Leschaux glaciers have not been coalescing (Gardent 2014).

A notable feature of Talèfre glacier is the presence of a triangle-shaped rock islet at its center, named “Jardin de Talèfre” (i.e. Talèfre garden). This islet is bounded by moraines (Fig. 2), that span 2640 to 3040 m a.s.l. (Jordan 2010). Because of its remoteness, this area has not been disturbed by human activity. According to the French État-Major maps (surveyed in 1820-1866) and first available photographs made by August-Rosalie Bisson in 1862 the “Jardin de Talèfre” was not ice-covered during the last LIA maximum (de Decker Heftler 2002).

Methodology

Moraine mapping

Several so-far undated moraine ridges preserved on the islet were studied and represented in the geomorphological map of Fig. 3. The mapping was conducted through field work and interpretation of aerial images from the year 2015 (French national institute of geographic and forest information - IGN, 50 cm resolution).

Sample collection

Fourteen samples were collected in the “Jardin de Talèfre”, 2 from smoothed bedrock surfaces resulting from glacial erosion and 12 from moraine boulders (Fig. 3). Bedrock samples were

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3 153 taken from the surface of the rocky islet, and attention was paid to avoid the possibility of cover
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5 154 by snow or vegetation and significant erosion by choosing sloping spots with glacial
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8 155 morphologies (Fig. 2C). Both surfaces were covered by lichen. Moraine boulder samples were
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10 156 taken from the top of the boulders, which were embedded in the crest or slope of the targeted
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12 157 moraines (Fig. 2D). If we assume uninterrupted surface exposure, the bedrock exposure ages
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14 158 indicate when the “Jardin de Talèfre” became ice-free for the last time, while the boulder
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16 159 exposure ages represent the timing of the glacier culmination and moraine stabilization (for
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18 160 detailed discussion see section “Oscillation of Talèfre glacier since the Younger Dryas”).
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20 161 Samples were collected using a chisel, a hammer and a cordless angle grinder, and topographic
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22 162 shielding was evaluated in the field using a clinometer.
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28 164 *Laboratory procedures*

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33 166 All samples were processed at CEREGE (Aix-en-Provence, France). Samples were crushed and
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35 167 sieved to collect the 250-500 μm fraction. Quartz was first concentrated by magnetic separation
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37 168 and then isolated by successive leaching in a $\text{H}_2\text{SiF}_6/\text{HCl}$ mixture or by froth flotation when the
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39 169 mineralogy allowed it. The obtained quartz fraction was leached at least two times in a 5% HF
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41 170 - 5% HNO_3 solution, followed by 24h in ultra-sonic baths in a 1% HF - 1% HNO_3 solution in
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43 171 order to remove any remaining feldspars and to clean the grains from atmospheric ^{10}Be . Purified
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45 172 quartz was completely dissolved in concentrated HF after addition of ~ 0.1 g of an in-house ^9Be
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47 173 carrier solution (3025 ± 9 ppm; Merchel *et al.* 2008). Beryllium was extracted by successive
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49 174 alkaline precipitations of $\text{Be}(\text{OH})_2$ alternated with separation on anion and cation columns.
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51 175 Samples were then oxidized at 700 $^\circ\text{C}$ for one hour and the final BeO mixed with Nb powder
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53 176 and loaded into nickel cathodes. Measurements of the $^{10}\text{Be}/^9\text{Be}$ ratios by accelerator mass
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55 177 spectrometry (AMS) were conducted at the French national AMS facility ASTER (Arnold *et*
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3 178 *al.* 2010). Samples were calibrated against in-house standard STD-11 with a $^{10}\text{Be}/^9\text{Be}$ ratio of
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5 179 $(1.191 \pm 0.013) \times 10^{-11}$ (Braucher *et al.* 2015) and a ^{10}Be half-life of $(1.387 \pm 0.0012) \times 10^6$ years
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8 180 (Chmeleff *et al.* 2010; Korschinek *et al.* 2010). Analytical uncertainties combine ASTER
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10 181 counting statistics and stability ($\sim 5\%$; Arnold *et al.* 2010) and machine blank correction. The
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12 182 number of atoms ^{10}Be in the samples calculated from the corresponding $^{10}\text{Be}/^9\text{Be}$ ratios was
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14 183 then corrected from that of the chemical blanks, which had $^{10}\text{Be}/^9\text{Be}$ ratios of $(3.38 \pm 0.49) \times 10^{-15}$
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16 184 and $(3.48 \pm 0.50) \times 10^{-15}$ (Table 1).
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21 186 *Age calculations*

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26 188 Surface exposure ages of the here presented new data were computed with the CREp online
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28 189 calculator (Martin *et al.* 2017) using the Lal-Stone time corrected scaling scheme Lm, the
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30 190 ERA40 atmospheric reanalyses and the Atmospheric ^{10}Be -based VDM for geomagnetic
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32 191 database (see Martin *et al.* 2017 and references therein). We applied the production rate of
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34 192 4.10 ± 0.10 atoms $^{10}\text{Be} \text{ g}^{-1} \text{ a}^{-1}$ established by Claude *et al.* (2014) as it is the only regional ^{10}Be
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36 193 production rate available for Alpine sites and it is identical to the global ^{10}Be production rate
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38 194 resulting from the ICE-D database linked to CREp (4.11 ± 0.19 atoms $^{10}\text{Be} \text{ g}^{-1} \text{ a}^{-1}$). The two other
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40 195 production rates often used for studies in the Alps, i.e. the northeast North America “NENA”
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42 196 production rate (Balco *et al.* 2009) and the “Arctic” production rate (Young *et al.* 2013), which
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44 197 is almost identical with the recent Scotland production rate (Putnam *et al.* 2019), would lead to
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46 198 ages older by less than 2% and 1%, respectively. For easy comparison, we present in Table 1
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48 199 the exposure ages calculated with both the regional production rate and the “Arctic” production
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50 200 rate. We did not apply a correction for snow shielding, as we do not have sufficient quantitative
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52 201 data on snowfall during the Holocene, and all sampled surfaces are in wind-exposed locations.
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54 202 We did not correct for potential erosion, as we were careful to sample surfaces with least
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possible erosion. Applying a correction for considerable snow cover assuming 50 cm of snow for six months (Chenet *et al.* 2016) would lead to ages older by 6% to 8%. Applying a correction for erosion corresponding to an erosion rate of 1 mm ka⁻¹ (André 2002) would result in ages older by less than 1%.

For the regional comparison of glacier records, previously published data from 18 study sites (see references in Fig. 6) were homogenized by re-calculating the individual and mean ages using the same procedures as above (Table S1). Boulders identified as outliers in the original studies were kept as outliers. Applying the χ^2 test (2σ) by Ward & Wilson (1978), moraine age populations were tested for outliers, which were also discarded as indicated in Table S1. Exceptions were made for the moraine M7a from the Rougnoux valley, M2 from Falgin cirque and the lower moraine from Steingletscher, where all boulders (not identified as outliers by authors) were kept for the mean age calculation even if they statistically mismatch, as we do not have any argument for discarding a specific sample from the age population. At the Rougnoux valley site, several moraine mean ages do not agree with the stratigraphic order (M8-9, M7b, M4-5 and M6b) but these ages do not statistically differ; therefore, we choose to include them all in the comparison. We eliminated from the comparison the moraine ages that are based on only one sample and that have standard deviations >10% to make the comparison as accurate as possible. Once this selection made, only the sites with at least one moraine of Holocene age (<11.7 ka) were kept for the comparison in order to exclude sites with moraines that exclusively correspond to earlier Lateglacial advances. Boxleitner *et al.* (2019) dated 34 moraine boulders of several glacier systems in Meiental (central Alps), but we deliberately did not consider the study in our test, because each moraine was dated with only one boulder. Applying a realistic snow cover correction to all sites would equally shift all moraine ages to slightly older dates and would not affect the discussion.

Results

Geomorphological description of the Talèfre site

Two massive lateral moraines, up to 50 m high and 1 km long, frame the islet of the “Jardin de Talèfre” (purple moraines, including T4, in Fig. 2B and Fig. 3). In the eastern part of the islet, three very smooth lateral moraine ridges (T1, T2 and T3) are visible in stratigraphically older location relative to the massive moraine T4 on the eastern margin. They are less than 2 m high and between 100 and 200 m long. The two outermost of these three moraine ridges (T1 and T2) bear several large boulders suitable for ^{10}Be dating, while the ridge T3 does not. These three ridges (T1 to T3) are represented in dark blue in Fig. 2B and Fig. 3. Moraine mapping shows that they are leaning northward against a 20 m high rock slope at an altitude of ~ 2850 m a.s.l.; to the south, they are crosscut by the T4 moraine. Between the western massive ridge and the present glacier margin, a small ridge (T5) is visible (pink moraine in Fig. 3). It has no equivalent on the eastern side and has been probably deposited during one of the short advances/stillstands that occurred at Alpine sites during the general glacier retreat of the 20th century (Schimmelpfennig *et al.* 2014; Braumann *et al.* 2020).

Moraine and bedrock exposure ages of Talèfre glacier

The two bedrock surfaces were sampled at the center of the northern part of the “Jardin de Talèfre”. Moraine boulder samples were taken from three ridges in the eastern part of the islet, i.e. from the two stratigraphically outermost (oldest), smooth moraines T1 and T2, and from the innermost (youngest), massive moraine T4. Surface exposure ages are shown in Fig. 3 and detailed in Table 1. Probability plots of moraine boulder ages and arithmetic mean ages are shown in Fig. 4. The consistency of individual ages within one moraine age populations were

tested for outliers with the χ^2 test (2σ) by Ward & Wilson (1978), using the analytical uncertainties of the ages. In the text and in figures, individual ages are given with the analytical uncertainties, to allow for internal comparison, and arithmetic mean ages are given with their standard deviations, while Table 1 shows both analytical uncertainties and total uncertainties (including the ^{10}Be rate uncertainty). See figure and table captions for details.

All ages are in agreement with their stratigraphic order. The two bedrock samples, JDT-16-1 and JDT-16-2, give indistinguishable exposure ages of 12.4 ± 0.4 and 12.3 ± 0.3 ka, respectively. The three boulder ages from the outermost moraine T1 range between 12.2 ± 0.4 and 11.6 ± 0.3 ka, with a mean age of 12.0 ± 0.4 ka. Three boulder ages from moraine T2 are very well clustered between 11.4 ± 0.3 and 11.1 ± 0.3 ka and give a mean age of 11.2 ± 0.2 ka after discarding JDT-16-20 as an outlier (10.2 ± 0.3 ka). This boulder is less than 1 m high. It might thus have been exhumed or covered for a while by a layer of sediment, leading to an underestimation of its exposure age. On the innermost moraine, four boulders yield ages of 0.18 ± 0.02 , 0.22 ± 0.03 , 0.26 ± 0.02 and 0.33 ± 0.03 ka and one boulder is dated at 2.12 ± 0.07 ka.

According to this new set of exposure ages we propose a reconstruction of the past extents of Talèfre glacier, corresponding to moraines T1 and T4 (Fig. 5). The mapping of the exterior glacier margins relies solely on remnants of undated moraines, assumed variations in relation to catchment slopes, and 19th century maps from French État-Major.

Exposure ages of other Alpine chronologies

The previously published 18 Alpine glacier chronologies that are considered here are characterized by sets of lateral and/or latero-frontal moraines originally dated between ~ 13.5 and ~ 9.5 ka. Of these 18 sites, 11 meet the quality criteria presented in the section “Methodology” and yield recalculated ^{10}Be exposure ages between 14.5 and 9.8 ka. The

278 moraine ages of the other sites are either older (Val Viola, Gesso della Barra valley, Romanche
279 valley, Great Aletsch glacier, Ferwall site) or show large standard deviations (Clarée valley).
280 Probability curves for all retained moraine mean ages in each site are plotted in Fig. 6 in
281 comparison with the Talèfre glacier moraine ages.

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283 Discussion

284 *Oscillations of Talèfre glacier since the Younger Dryas*

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286 According to the exposure ages of the two bedrock samples, the “Jardin de Talèfre” deglaciation
287 started at ~12.4 ka, i.e. during the YD period, if a continuous exposure scenario is assumed.

288 Alternatively, the duration of ~12.4 ka could be composed of several exposure periods, e.g. a
289 few hundred years of ice-free conditions during the warmer Bølling/Allerød (~14.7 – 12.9 ka;
290 Lotter *et al.* 2000; Rasmussen *et al.* 2006) and subsequent burial by the re-advancing glacier at
291 the beginning of the YD. However, the only slightly younger mean age of moraine T1 (12.0 ± 0.4
292 ka) implies that “Jardin de Talèfre” was deglaciated during the YD and the central rock islet
293 has remained ice-free since that time. The mean ages of moraines T1 and T2 are
294 stratigraphically consistent (12.0 ± 0.4 and 11.2 ± 0.2 ka) and coincide with the YD/EH transition.
295 We propose that the T3 moraine was deposited during the EH, as its morphology is very similar
296 to T1 and T2 moraines. This suggests that the glacier underwent at least three advances or
297 stillstands after the beginning of its retreat ~12.4 ka ago until the EH.

298 Four out of the five ages acquired from moraine T4 span from 0.18 ± 0.02 to 0.33 ± 0.03
299 ka, with a mean age of 0.25 ± 0.06 ka. These results indicate that the moraine was last overtopped
300 at the end of the LIA, when several closely-spaced advances occurred. One out of the five
301 samples of T4 is significantly older, 2.12 ± 0.07 ka (JDT-16-3). This older age can be either
302 explained by isotope inheritance from a previous exposure duration or by accretion phases of

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3 303 this massive moraine that pre-dated the deposition of the LIA boulders. Such advances prior to
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5 304 the LIA have been reported for other glaciers in the French Alps: between ~3.5 and 0.7 ka at
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8 305 Mer de Glace (Le Roy *et al.* 2015) and ~2.1 ka at Argentire glacier (Le Roy 2012) and in the
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10 306 Ecrins-Pelvoux massif (Le Roy *et al.* 2017). However, our data do not allow us to decide which
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13 307 of the two hypotheses is more likely.

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15 308 Fig. 5 shows the large YD/EH and LIA extents of Talfre glacier in comparison to that
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17 309 of 2008. During the YD/EH and the LIA, the glacier is still connected with the Mer de Glace,
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19 310 and the LIA extent is only slightly smaller than that of the YD/EH, while in 2008 the glacier
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22 311 size is significantly reduced.

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24 312 We do not have any evidence of glacier advances between the EH and the Late
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26 313 Holocene, which is consistent with a warmer and dryer climate, and thus unfavorable for glacier
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28 314 growth in the Alps, during this period (Ivy-Ochs *et al.* 2009). Therefore, we assume that Talfre
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31 315 glacier was smaller than its LIA extent during this period.

32 33 316 34 35 317 *Comparison with other Alpine glacier chronologies*

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40 319 The YD/Holocene chronology of Talfre glacier is very similar to the one recently established
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42 320 at the nearby Argentire glacier (Protein *et al.* 2019). A similar set of bedrock surfaces (~2450
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44 321 m a.s.l.) and five lateral moraines (~2400 - 2150 m a.s.l.), dated by *in situ* cosmogenic ¹⁰Be to
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47 322 between ~11.7 and ~10.4 ka, provide evidence of deglaciation at the end of the YD and at least
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49 323 five glacier re-advances or stillstands during the YD/EH transition. When comparing these two
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52 324 age sets (Fig. 6D, E), it appears that the ages of the two outermost moraines of both glaciers are
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54 325 undistinguishable within uncertainties, and that both glaciers oscillated several times during the
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56 326 YD/EH transition. In this comparison, it must be kept in mind that T3 moraine of Talfre glacier
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58
59 327 could not be dated but seems to be only slightly younger than T1 and T2 moraines (see section
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“Moraine and bedrock exposure ages of Talèfre glacier”) and might therefore be similar to the innermost EH ridge of Argentièrè glacier dated to 10.4 ± 0.2 ka. These moraines of Talèfre and Argentièrè glaciers thus probably represent quasi-synchronous, successive glacier stillstands within the dynamic YD/EH retreat from their larger early-YD extents.

Moraines at multiple other locations across the Alps have previously been assigned to the YD/EH transition based on ^{10}Be dating. Eleven of them were selected here, based on their data quality (see section “Exposure ages of other Alpine chronologies”), together with our new Talèfre chronology for this regional compilation. They are geographically distributed across the European Alps (France, Switzerland, Italy and Austria), situated between longitudes $\sim 6.5^\circ\text{E}$ and 10°E and latitudes 45.5°N and 47°N (Fig. 6). All of these twelve moraine chronologies indicate deposition of at least one moraine between ~ 12 and ~ 10 ka ago, i.e. during the YD/EH transition (Fig. 6). Further, at most sites, there is evidence of deposition of multiple lateral and/or latero-frontal moraines during this period. Successive moraines are located quite close to each other and in proximity to the LIA moraines, suggesting only slightly smaller glacier extents during the LIA than during the YD/EH compared to today’s reduced glacier sizes like at Talèfre glacier (Fig. 5). The highest spatial spread is found at Arsine glacier displaying YD/EH frontal moraines that are located about 1 km downstream of the LIA moraines (Schimmelpfennig *et al.* 2019). At several other sites in the Alps, lateral YD/EH moraines exist within <300 m-distance from lateral LIA moraines (Schimmelpfennig *et al.* 2012, 2014; Moran *et al.* 2016a; Hofmann 2018).

Thus, based on the consistency of the ^{10}Be cosmogenic exposure ages, we show here that millennial-scale synchronous glacier fluctuations occurred in the European Alps, and that glaciers oscillated multiple times at centennial or sub-centennial scale superposed on a general retreat trend during the 12-10 ka period. However, based on the available datasets it is not possible to temporally correlate specific moraines between sites or to give evidence of a decadal

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or centennial-scale synchronicity of glacier dynamics within the ~12-10 ka period, because of (i) the analytical and geomorphic uncertainties related to the dating method; (ii) the probably incomplete preservation of the moraine records or no deposition of the moraine; and (iii) undated ridges within the investigated moraine sequences at several sites.

For the sites where ELA depressions between the YD/EH transition and the LIA have been calculated using the Accumulation Area Ratio (AAR) method with a ratio of 0.67 (Table S1), no significant correlations between moraine ages and ELA depressions are observed (Fig. S1). In fact, the large range of ELA depressions supports the findings by Boxleitner *et al.* (2019) that indicate ELA depressions should not be used to temporally correlate moraines from different valleys, as previously proposed by Maisch (1981). This reflects the fact that ELA variations of individual glaciers might be significantly influenced by local precipitation amounts and local glacier morphology, as well as the limitations of the AAR method (e.g. Reitner *et al.* 2016; Boxleitner *et al.* 2019).

Link of the Alpine glacier oscillations with regional climate

Analysis of the geographical characteristics of the 12 selected sites implies that their millennial synchronicity is independent of their altitude range (at present between 1500 and 3400 m a.s.l.), orientation (all directions) and size (0.3 to 14 km²) (Fig. S1). In the absence of common topographic characteristics of the glaciers, the only process to explain the millennial-scale synchronous fluctuations of independent ice mass fluctuations across the European Alps is a regional climatic signal that prevailed over local factors. Therefore, we compare here the moraine chronologies with regional - and high-latitude - independent climate proxy records in an effort to elucidate the spatial reach of the climatic evolution underlying the quasi-simultaneous glacier oscillations across the European Alps within their dynamic retreat.

Glaciers are sensitive to a variety of climatic factors, but the main driver for large-scale, long-term glacier fluctuations in the mid- and high latitudes is the impact of summer temperature variations (e.g. Oerlemans 2005; Solomina *et al.* 2015) on which we therefore focus. In addition, over the last century, annual mass balance of Alpine glaciers was mainly driven by summer melting (Huss *et al.* 2010). In the Alpine and pre-Alpine region, evidence of a temperature minimum in the middle of the YD and several distinct cold events during a general warming trend within the 12-10 ka period comes from various high-resolution palaeorecords that reflect variations either in July air temperatures (based on chironomid-inferred records from lakes in eastern France and the northern and central Swiss Alps; Heiri *et al.* 2015; Fig. 6A(3)) or in mean annual temperatures, based on stable isotope measurements in stalagmites (Wurth *et al.* 2004; Affolter *et al.* 2019; Fig. 6A(2)) and biogenically precipitated carbonates in lake sediments (Schwander *et al.* 2000). These decadal to centennial cold peaks seem to have amplitudes in the order of ~ 0.5 °C, i.e. similar to the multidecadal fluctuations that drove the succession of LIA moraines in the Alps (Oerlemans 2005). Regional cold peaks could therefore be sufficient to lead to the YD/EH succession of glacial culminations or still-stands during the general deglaciation observed in the Alps. In this case, it is most likely that decadal to centennial glacier oscillations occurred quasi-simultaneously across the Alps.

Link of the Alpine glacier extents with seasonality

It is remarkable that several glaciers around 12 ka ago, i.e. at the end of the YD, were only slightly larger than during the EH. This is the case for Rougnoux valley, Arsine, Talèfre, Argentièrre, Tsidjiore Nouve, Belalp and Falgin Cirque glaciers, where the ages of the outermost moraine fall into the YD or their uncertainties overlap with it (green probability curves in Fig. 6) and the ages of the innermost moraine fall into the EH. This small difference between late

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YD and EH glacier extents seems inconsistent with the drastic YD temperature cooling and abrupt mean annual temperature increase that characterizes the boundary between these two periods (Fig. 6A). A linear relationship between oxygen isotope record and temperature would suggest that the coldest YD temperatures of the Greenland sites were only slightly higher than those of the Last Glacial Maximum (e.g. Rasmussen *et al.* 2008) when Alpine glaciers reached until the Alpine foreland (Ivy-Ochs *et al.* 2008). Much larger glacier extents might therefore be expected during and at the end of the ~1 ka lasting YD, especially compared to the EH and LIA. In addition, the ¹⁰Be bedrock ages from outboard of the oldest dated moraines of Talèfre and Argentièrè glaciers suggest that these glaciers have been retreating prior to the deposition of their respective late YD moraine. These results are in line with ¹⁰Be-based glacier chronologies from Greenland, also providing evidence of ice mass downwasting throughout the YD, which lasted until the EH (Rainsley *et al.* 2018 and references therein).

The apparent mismatch between the glacier proxy records and the mean annual temperature climate proxies could be due to a difference in seasonality between the YD and the EH. This concept of seasonality was first proposed by Denton *et al.* (2005) to explain the discord between the drastic change of the mean annual temperature observed in Greenland ice cores and the slight snowline change of glaciers in Greenland and Europe during the YD and EH. The YD, as well as other abrupt climate changes, has been qualified as a winter phenomenon (Denton *et al.* 2005), meaning that the considerable annual temperature switches observed in palaeoclimate records from Europe and Greenland are mainly due to changes in winter temperature, while summer temperature did only undergo minor changes (Renssen & Isarin 2001; Buizert *et al.* 2014; Schenk *et al.* 2018).

In the Alpine region, a decrease of the seasonality between the YD and EH is supported by the difference of mean annual and summer warming estimated from the above-mentioned recent independent proxy studies (Heiri *et al.* 2015; Affolter *et al.* 2019; Fig. 6A(2, 3)). The

stable isotope measurements in speleothems suggest an increase of the mean annual temperature of ~ 5.4 °C at the YD/EH transition (Affolter *et al.* 2019; Fig. 6A(2)), while a summer temperature increase of only ~ 2.7 °C is inferred from the chironomid-based lake records (Heiri *et al.* 2015; Fig. 6A(3)). This difference between annual and summer temperature could be explained by a greater winter warming of more than 7 °C if winter and summer periods are assumed of equal duration. Furthermore, while the mean temperature increase in Greenland occurred in less than one hundred years (Rasmussen *et al.* 2006; Fig. 6A(1)), the European records indicate a more gradual warming distributed over several hundred years (Fig. 6A(2, 3)). Our results support this concept, i.e. relatively mild YD summer temperatures could explain why the glaciers in the Alps did not undergo a much more pronounced retreat at the YD/EH transition, but glaciers retreated gradually from their only slightly larger YD extents.

Link of the Alpine glacier oscillations with high latitude climate

The clear correlation of high-resolution proxy records from the Alps and the Alpine foreland (Heiri *et al.* 2015; Affolter *et al.* 2019; Fig. 6A(2, 3)) with the pattern of oxygen isotope variations in Greenland ice cores (e.g. Rasmussen *et al.* 2006; Fig. 6A(1)) strongly suggests a climatic link of the Alpine region with the northern high latitudes. In this context, a recent study by Young *et al.* (2020) provides evidence of a succession of re-advances or stillstands of mountain glaciers, the Laurentide Ice Sheet and Greenland Ice Sheet in the Baffin Bay region at ~ 11.8 - 11.6 , ~ 10.4 and ~ 9.3 ka. The striking similarity of these moraine-based chronologies from and near western Greenland with those from the European Alps provide further evidence of a teleconnection between both regions at that time and that the observed glacier fluctuations were due to a common climate evolution on extra-regional, maybe even hemispheric scale.

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The question arises which large-scale climate mechanisms led to this teleconnection and the contrasted seasonality imprinted in the above-discussed summer temperature and mean annual temperature records. The abrupt onset of the YD cooling and the cold peaks of the Late YD and EH in the high latitudes are thought to have been triggered by freshwater input from the Laurentide Ice Sheet into the North Atlantic (Jennings *et al.* 2015) leading to a reduction of the Atlantic Meridional Ocean Circulation (AMOC) (Renssen *et al.* 2015). The related development of expanded wintertime sea ice in the North Atlantic shifted the westerlies south leading to wind-transported cool temperatures to Europe (Brauer *et al.* 2008; Bakke *et al.* 2009), and therefore to the Alps. During the second half of the YD, brief sea-ice breakups were led by incursions of the Atlantic circulation and were associated with a northwards migration of the polar front and westerly storm tracks (Bakke *et al.* 2009) until the system abruptly switched to the interglacial state at the onset of the Holocene (Rasmussen *et al.* 2006). The signature of these climatic mechanisms is markedly imprinted in the abrupt variations of the mean annual temperature proxy records, but to a lesser degree in the glacier mass balance fluctuations. Robust evidence of drastic glacier expansion in the Alps and the high latitudes in relation with the strong annual cooling observed in the first part of the YD is rare so far. Early to mid-YD moraines are preserved at a few sites in the Alps and at other sites in the North Atlantic region (e.g. Belalp and Aletsch Glacier, Kelly *et al.* 2004; Schindewolf *et al.* 2012; Julier Pass, Ivy-Ochs 2015; Alaska, Young *et al.* 2019; Norway, Wittmeier *et al.* 2020), but the corresponding glacier extents were relatively modest compared to significant annual temperature decrease. In addition, the deglaciation during the YD/EH transition was rather gradual and punctuated by closely-spaced short-lived re-advances or stillstands. This could be explained by several mechanisms. Sea ice expansion occurs predominantly in winter and might therefore not directly result in a summer temperature decrease. In addition, the drastic YD cooling recorded by the mean annual temperature proxies was enhanced by an increased strength of wind only during

autumn, winter and spring, which transported cold air from the ice-covered North Atlantic region and led to cool and dry winters in Europe (Brauer *et al.* 2008). This phenomenon did not occur during summer due to atmospheric blocking of the cold westerlies over Fennoscandia leading to warm summers during the YD (Schenk *et al.* 2018). Therefore, and because snowfall was reduced during the dry winters of the cold periods, greater glacier expansion did not occur at many sites. Nevertheless, summers were very short, probably because winter sea ice expansion lasted for a long period of the year, leading to a reduced glacier melt season (Young *et al.* 2020). This might explain why glaciers responded with successive glacier advances or stillstands to these dynamic changes in seasonality both in high latitudes and Europe.

Conclusions

Our new chronology based on 14^{10}Be cosmogenic exposure ages on moraine and bedrock from the “Jardin de Talèfre” suggests that the central rock islet of the “Jardin de Talèfre” was deglaciated during the YD and that Talèfre glacier retreated from its YD position starting ~12.4 ka ago. This retreat was interrupted by several episodes of stillstands before a major retreat during the EH. No evidence of re-advances before the Late Holocene is recorded and the most remarkable advance occurred during the LIA, which was only slightly less extensive than the late YD and EH glacier extents at our high-elevation site. This behaviour is very similar to that of the nearby Argentièrè glacier. The comparison of the YD/EH transition moraine chronology of Talèfre glacier with those of 11 other Alpine glaciers scattered throughout the Alps suggests a period of roughly synchronous glacial behaviour across the Alps between 12 and 10 ka ago. Several glaciers deposited multiple moraines, which attest to regional-scale stillstands or re-advances during the deglaciation. Similarities with palaeotemperature records from the Alpine region as well as with glacial culminations and temperature records from Greenland strongly

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suggest not only a regional common climate driver, but also a teleconnection with the high latitudes, thus reinforcing the hypothesis postulated in earlier studies of a link between YD and EH Alpine glacier fluctuations and Greenland climate change. However, the relatively small amplitude of the YD glacier extents compared to the Holocene glacier extents is inconsistent with the large and abrupt temperature shifts imprinted in mean annual proxy records in both regions. We suggest that this inconsistency is due to a pronounced change in seasonality, with YD winters cold and dry and YD summers only slightly cooler than those of the EH. This seasonality difference controlled YD glacier extents and led to YD glaciers only slightly larger than during the EH.

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528 *Data availability.* - The data that support the findings of this study are available from the
529 corresponding author upon reasonable request.

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Legends

Fig. 1. Location of Talèfre glacier and study area: the “Jardin de Talèfre”. A. Location of the Mont Blanc massif (red dot) in the Alps (red dotted line). B. The Mont Blanc massif and present-day glacier extents on a Landsat 8 image (April 2015); French glaciers: Gardent *et al.* (2014); Swiss glaciers: Fischer *et al.* (2014); Italian glaciers: Smiraglia *et al.* (2015). Talèfre glacier is highlighted in dark blue, the “Jardin de Talèfre” is located in the red square.

Fig. 2. A. Photograph of the “Jardin de Talèfre” in 1890 (source: "Les Amis du Vieux Chamonix"). B. Oblique aerial imagery of the “Jardin de Talèfre” in September 2016. Talèfre glacier is highlighted in light blue, sampled moraines T1 and T2 (dark blue) and Holocene moraines, including T4 (purple) are shown. Bedrock sample locations are shown as red dots and some of the sampled boulders as green dots (photograph by J.-F. Buoncristiani). C. Example of sampled bedrock surface. D. Example of sampled boulder on T2 moraine.

Fig. 3. Moraines and sample locations in the “Jardin de Talèfre”, mapped on the 5 m IGN DEM RGE ALTI and aerial images. Present-day glacier extent (in blue) is from Gardent *et al.* (2014). Dashed lines represent undated moraines, the age of which is assumed according to their relative position and morphology. Individual surface exposure ages are listed in the white boxes with their analytical uncertainties and mean ages of the landforms in the colored bands (arithmetic means and standard deviations). Outliers are in italic grey font.

Fig. 4. Probability plots of the individual moraine sample ages (black curves, analytical uncertainty only) from Talèfre glacier. Colored curves, vertical lines and bands represent the

819 summed probability plots, arithmetic means and standard deviations, respectively,
820 corresponding to the age labeled in each box, only calculated for T1 and T2.

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822 Fig. 5. Reconstruction of past extents of Talèfre glacier, based on mapping of the preserved
823 moraines and interpretation of ^{10}Be dating, for the Younger Dryas/Early Holocene transition
824 (A) (12 ka, stage T1 moraine), LIA maximum (B) and 2008 CE (C) (Gardent *et al.* 2014).
825 Dashed glacier limits correspond to hypothetical extents.

827 Fig. 6. Chronology of culminations of Talèfre glacier based on the ^{10}Be mean ages of the glacial
828 landforms in this study (D) compared to independent palaeoclimate records (A) and to other
829 glacier chronology studies in the Alps also using ^{10}Be moraine dating (B-M). Each moraine is
830 represented by a summed probability curve along with the number of dated boulders ($n = x$),
831 vertical lines and colored band represent the arithmetic means and standard deviations (1σ),
832 colors represent the stratigraphic order of the moraines (from outer to inner: green, blue, purple,
833 pink). A(1). Oxygen isotope record in Greenland ice core NGRIP (Rasmussen *et al.* 2006;
834 Vinther *et al.* 2006), blue bands represent the Younger Dryas and cold peaks during the Early
835 Holocene as identified in the Greenland ice core record. A(2). Oxygen isotope record from
836 speleotem fluid inclusion from Milandre cave, Switzerland (Affolter *et al.* 2019). A(3). Stacked
837 chironomid-inferred July air temperature from the Alps (Heiri *et al.* 2015), black arrows point
838 to peaks mentioned in the main text. Alpine glacier sites in panels B, C and E to M are presented
839 from east to west with their locations shown by red dots on the map. B. Rognoux valley
840 (Hofmann 2018). C. Arsine glacier (Schimmelpfennig *et al.* 2019). E. Argentièr glacier (Protin
841 *et al.* 2019). F. Tsidjiore Nouve glacier (Schimmelpfennig *et al.* 2012). G. Belalp cirque
842 (Schindelwig *et al.* 2012). H. Steingletscher (Schimmelpfennig *et al.* 2014). I. Kromer valley
843 (Kerschner *et al.* 2006; Moran *et al.* 2016b). J. Kloster valley (Moran *et al.* 2016b). K. Falgin

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glacier (Moran *et al.* 2016a). L. La Mare glacier (Baroni *et al.* 2017). M. Careser glacier (Baroni *et al.* 2017). Vertical dashed lines delimit the YD/EH transition period between 12 and 10 ka.

Table 1. Characteristics and ¹⁰Be measurement details of samples and blanks measured in this study. Outliers are highlighted by (O). Results of the exposure ages are presented using both the regional production rate (Claude *et al.* 2014) and the “Arctic” production rate (Young *et al.* 2013). Exposure ages are indicated with external uncertainties and internal uncertainties (including the ¹⁰Be production rate uncertainty) in parentheses. Mean landform ages of T1 and T2 are indicated with standard deviations and total uncertainties (including the ¹⁰Be production rate uncertainty) in parentheses.

Supporting Information

Fig. S1. Geographic characteristics of the Alpine glaciers selected for the comparison and presented in Fig. 6 and comparison of ELA depression (relative to LIA). A. Wind rose indicating the orientation of each glacier. B. Graph representing the present-day length of the glaciers versus their maximum and minimum elevation. C. ELA depression (relative to LIA, calculated with the AAR methods and 0.67 ratio) versus the age of the studied stadial. Letters correspond to panels and locations in Fig. 6. As Rougnoux glacier (Hofmann 2018) has now disappeared, it is not represented in A and B.

Table S1. ¹⁰Be exposure ages from the previously published moraine records in the Alps that were considered in the comparison with the moraine ages from Talèfre glacier and associated ELA depressions (relative to LIA) when calculated in the original study. For comparison, all

ages were calculated using the same method as applied for the Talèfre samples (see “Age calculations”). Outliers (O) were determined either based on original authors choice or on the χ^2 test (2σ) as applied to the ages from Talèfre moraines. For samples from Cossart *et al.* (2012), a sample thickness of 2 cm was assumed as this information is not given in the original paper. When not specified by authors, we assumed a rock density of 2.7 g cm⁻³. Sites or moraines that do not fulfill the selection criteria detailed in the main text are in italic while sites that met the selection criteria for comparison are in bold. Capital letters in parentheses correspond to panels and location in Fig. 6. ELA are calculated with the AAR method and a 0.67 ratio.

Table 1. Data for the 10Be/9Be ratios and ages of the moraine samples														
Sample name	Latitude (dd)	Longitude (dd)	Elevation (m a.s.l.)	Thickness (mm)	Shielding factor	Quartz weight (g)	Carrier (mg 9Be)	Associated blank	10Be/9Be x10-14	(10Be) (x104 at g-1)	Regional PR		Arctic PR	
											10Be age (ka)	1σ uncertainties (ka)	10Be age (ka)	1σ uncertainties (ka)
Moraine samples														
T1											11.97±0.36 (0.46)		12.07±0.37 (0.83)	
JDT-16-17	45.91642	6.98233	2825	45	0.9063	24.67	0.3053	BLC-15Jan18	44.9±1.4	36.8±1.2	12.14	0.37 (0.47)	12.24	0.37 (0.82)
JDT-16-13	45.91604	6.98249	2817	27	0.9635	15.35	0.3072	BLC-15Jan18	30.13±0.93	39.8±1.2	12.21	0.36 (0.46)	12.32	0.37 (0.83)
JDT-16-16	45.91564	6.98224	2798	22	0.9635	28.65	0.3078	BLC-15Jan18	52.3±1.6	37.3±1.2	11.56	0.34 (0.42)	11.65	0.35 (0.77)
T2											11.21±0.17 (0.32)		11.30±0.18 (0.71)	
JDT-16-20 (O)	45.91667	6.98301	2835	42	0.9638	26.58	0.3071	BLC-15Jan18	42.7±1.4	32.7±1.1	10.15	0.32 (0.40)	10.24	0.32 (0.68)
JDT-16-19	45.91647	6.98318	2827	25	0.9638	21.49	0.3070	BLC-15Jan18	39.5±1.3	37.4±1.2	11.40	0.34 (0.43)	11.50	0.35 (0.76)
JDT-16-22	45.91636	6.98320	2824	24	0.9638	32.44	0.3069	BLC-15Jan18	57.6±1.8	36.2±1.1	11.07	0.33 (0.41)	11.16	0.33 (0.73)
JDT-16-18	45.91599	6.98324	2805	35	0.9527	20.18	0.3083	BLC-15Jan18	34.9±1.2	35.3±1.2	11.15	0.35 (0.44)	11.24	0.36 (0.75)
T3														
JDT-16-3 (O)	45.91918	6.98241	2951	20	0.9677	37.83	0.3048	BLC-08Dec17	13.43±0.42	7.05±0.23	2.12	0.07 (0.09)	2.15	0.07 (0.19)
JDT-16-9	45.91840	6.98260	2923	38	0.9625	31.65	0.3033	BLC-08Dec17	1.33±0.11	0.64±0.08	0.18	0.02 (0.02)	0.18	0.02 (0.02)
JDT-16-10	45.91788	6.98305	2903	28	0.9625	31.26	0.3017	BLC-08Dec17	1.56±0.11	0.78±0.08	0.22	0.02 (0.02)	0.22	0.03 (0.03)
JDT-16-11	45.91683	6.98430	2842	26	0.9434	36.61	0.3063	BLC-08Dec17	1.88±1.20	0.86±0.07	0.26	0.03 (0.03)	0.26	0.02 (0.03)
JDT-16-21	45.91583	6.98377	2796	22	0.9623	33.60	0.3010	BLC-08Dec17	2.16±0.12	1.09±0.08	0.33	0.03 (0.03)	0.33	0.03 (0.03)
Bedrock samples														
JDT-16-1	45.91873	6.98128	2950	32	0.9667	20.09	0.3015	BLC-08Dec17	44.2±1.4	44.0±1.4	12.43	0.38 (0.48)	12.54	0.39 (0.85)
JDT-16-2	45.91894	6.98219	2943	26	0.9677	17.83	0.3048	BLC-08Dec17	38.6±1.1	43.7±1.2	12.31	0.33 (0.44)	12.42	0.34 (0.81)
Blanks														
Blank name	Carrier (mg 9Be)		10Be/9Be x10-14		Number of 10Be atoms (x104 at)									
BLC-8Dec17	0.3040		0.338±0.049		6.9±1.0									
BLC-15Jan18	0.3075		0.348±0.050		7.2±1.0

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Moraine and sample name	Regional PR			Arctic PR			Δ ELA compare to LIA (m)
	Recalculated ¹⁰ Be age (ka)	1 σ uncertainty (ka)	Moraine age (ka)	Recalculated ¹⁰ Be age (ka)	1 σ uncertainty (ka)	Moraine age (ka)	
Rognoux Valley (B) (Hofmann 2018)							
VdR M8d			12.3±0.2			12.4±0.2	-220
VdR-26	12.43	0.42 (0.51)		12.53	0.42 (0.86)		
VdR-27	12.34	0.42 (0.50)		12.44	0.42 (0.86)		
VdR-28	12.01	0.53 (0.59)		12.11	0.53 (0.90)		
VdR M7a			11.3±0.9			11.3±0.9	
VdR-23	10.38	0.42 (0.47)		10.45	0.42 (0.74)		
VdR-24	12.25	0.56 (0.62)		12.35	0.56 (0.93)		
VdR-25-1	11.76	0.40 (0.48)		11.85	0.40 (0.81)		
VdR-25-2	10.66	0.32 (0.40)		10.74	0.32 (0.70)		
VdR M4-5b			11.5±0.5			11.6±0.5	
VdR-10	11.17	0.37 (0.45)		11.27	0.37 (0.75)		
VdR-11	11.83	0.92 (0.95)		11.93	0.92 (1.16)		
VdR-12 (O)	7.41	0.38 (0.41)		7.46	0.38 (0.56)		
VdR M8a			11.2±0.6			11.3±0.6	
VdR-16	11.74	0.38 (0.47)		11.83	0.38 (0.80)		
VdR-17	10.64	0.40 (0.46)		10.71	0.40 (0.74)		
VdR-18	11.21	1.19 (1.21)		11.30	1.19 (1.36)		
VdR M6b			11.0±0.5			11.1±0.5	
VdR-7	11.42	0.40 (0.48)		11.50	0.40 (0.78)		
VdR-8	11.17	0.66 (0.70)		11.25	0.66 (0.93)		
VdR-9	10.53	0.36 (0.42)		10.62	0.36 (0.71)		
VdR M6a			10.9±0.4			11.0±0.4	
VdR-13	1.38	0.12 (0.13)		1.39	0.12 (0.16)		
VdR-14	11.12	0.34 (0.42)		11.21	0.34 (0.74)		
VdR-15	10.62	0.38 (0.45)		10.69	0.38 (0.73)		
VdR M7b			10.9±0.2			11.0±0.2	
VdR-4	10.66	0.36 (0.43)		10.74	0.36 (0.73)		
VdR-5	11.13	0.55 (0.60)		11.22	0.55 (0.86)		
VdR-6	10.83	0.33 (0.41)		10.92	0.33 (0.71)		
VdR M8-9			10.4±0.5			10.5±0.5	
VdR-1	10.97	0.35 (0.43)		11.06	0.35 (0.73)		
VdR-2	10.35	0.43 (0.49)		10.43	0.43 (0.75)		
VdR-3	10.00	0.36 (0.42)		10.07	0.36 (0.69)		
Arsine Glacier (C) (Schimmelpfennig <i>et al.</i> 2019)							
Frontal moraine			11.4±0.4				-220
Petit Tabuc lateral moraine			11.0±0.3			11.1±0.3	
Rif de la Planche frontale moraine			10.9±0.3			11.0±0.3	-164
Right lateral complex			10.4±0.6			10.5±0.6	
Piano del Praiet (Federici <i>et al.</i> 2008)							
PDP moraine			14.5±0.9			14.6±0.9	
PDP1	15.27	0.61 (1.32)		15.42	0.61 (1.08)		
PDP2	15.00	0.83 (1.44)		15.16	0.83 (1.21)		
PDP3	14.35	0.58 (1.24)		14.52	0.58 (1.02)		
PDP4	13.03	0.59 (1.18)		13.45	0.59 (0.99)		
Argentière Glacier (E) (Protin <i>et al.</i> 2019)							
L1			11.7±0.7			10.8±0.7	-105
ARG-16-9	12.21	0.58 (0.64)		12.31	0.58 (0.94)		
ARG-16-10 (O)	10.10	0.65 (0.68)		10.18	0.65 (0.88)		
ARG-16-11	10.63	0.42 (0.48)		10.70	0.42 (0.75)		
ARG-16-12	11.88	0.68 (0.73)		11.97	0.68 (0.99)		
ARG-16-13	11.93	0.55 (0.61)		12.03	0.55 (0.91)		

L2				10.4±0.2			10.5±0.2
	ARG-16-1	10.28	0.66 (0.70)		10.36	0.66 (0.90)	
	ARG-16-2	10.39	0.66 (0.70)		10.47	0.66 (0.90)	
	ARG-15-11	10.65	0.32 (0.40)		10.73	0.32 (0.70)	
L3				10.4±0.4			10.5±0.4
	ARG-16-3	10.61	0.33 (0.41)		10.68	0.33 (0.70)	
	ARG-16-4	10.52	0.32 (0.40)		10.60	0.32 (0.70)	
	ARG-15-12	10.87	0.33 (0.40)		10.96	0.33 (0.72)	
	ARG-16-5 (O)	17.42	1.51 (1.55)		17.56	1.51 (1.82)	
	ARG-16-6	10.04	0.30 (0.38)		10.11	0.31 (0.67)	
	ARG-16-7	10.01	0.34 (0.41)		10.10	0.34 (0.69)	
<i>Romanche Valley (Chenet et al. 2016)</i>							
M7				12.5±0.8			12.5±0.9
	RO-20	13.04	0.67 (0.73)		13.16	0.67 (1.04)	
	RO-21	11.86	0.58 (0.63)		11.93	0.58 (0.92)	
	RO-22 (O)	16.15	1.86 (1.88)		16.27	1.86 (2.09)	
	RO-23 (O)	7.72	0.61 (0.64)		7.79	0.61 (0.76)	
	RO-24 (O)	12.98	3.62 (3.61)		13.09	3.62 (3.71)	
M6				11.9±0.8			12.0±0.8
	RO-15	12.29	0.93 (0.96)		12.38	0.93 (1.19)	
	RO-16	10.30	0.89 (0.92)		10.39	0.89 (1.08)	
	RO-17	11.72	0.63 (0.68)		11.81	0.63 (0.95)	
	RO-18	13.10	0.69 (0.76)		13.23	0.69 (1.05)	
	RO-19 (O)	2.08	0.41 (0.41)		2.1	0.41 (0.43)	
<i>Clarée Valley (Cossart et al. 2012)</i>							
Stage 2 moraine				12.7±1.1			12.8±1.1
	CLA_08_13a	11.43	0.85 (0.88)		11.51	0.85 (1.09)	
	CLA_08_13b	13.21	0.64 (0.71)		13.33	0.64 (1.02)	
	CLA_08_14	13.51	0.93 (0.98)		13.61	0.93 (1.23)	
Stage 3 moraine				11.4±1.2			11.5±1.2
	CLA_08_01	12.15	0.42 (0.50)		12.25	0.42 (0.84)	
	CLA_08_02a	12.63	0.74 (0.80)		12.74	0.74 (1.06)	
	CLA_08_02b (O)	14.09	0.69 (0.76)		14.23	0.69 (1.08)	
	CLA_08_03	11.78	0.54 (0.59)		11.87	0.54 (0.89)	
	CLA_08_04	10.74	0.61 (0.65)		10.82	0.61 (0.88)	
	CLA_08_05	12.51	0.81 (0.85)		12.60	0.81 (1.10)	
	CLA_08_06	10.12	3.19 (3.17)		10.20	3.19 (3.24)	
	CLA_08_10	9.83	3.79 (3.76)		9.92	3.79 (3.83)	
<i>Tsidjiore Nouve Galiier (F) (Schimmelpfennig et al. 2012)</i>							
Crest				11.7±0.2			11.8±0.2
	ARO-4	11.94	0.18 (0.33)		12.04	0.18 (0.74)	
	ARO-1	11.90	0.21 (0.35)		12.01	0.21 (0.75)	
	ARO-6	11.87	0.24 (0.36)		11.98	0.24 (0.76)	
	ARO-55	11.63	0.18 (0.32)		11.73	0.18 (0.72)	
	ARO-5	11.58	0.21 (0.33)		11.67	0.21 (0.72)	
	ARO-52	11.39	0.20 (0.33)		11.50	0.20 (0.70)	
	ARO-59	11.34	0.20 (0.33)		11.44	0.20 (0.70)	
	ARO-3 (O)	11.20	0.19 (0.32)		11.29	0.19 (0.69)	
	ARO-56 (O)	11.05	0.19 (0.32)		11.14	0.19 (0.68)	
	ARO-2 (O)	10.96	0.19 (0.31)		11.05	0.19 (0.67)	
Recessional ridges				11.4±0.2			11.5±0.2
	ARO-21_2010Apr	11.72	0.28 (0.38)		11.82	0.28 (0.75)	
	ARO-21	11.62	0.23 (0.34)		11.72	0.23 (0.73)	
	ARO-22	11.48	0.26 (0.37)		11.58	0.26 (0.73)	
	ARO-21-re	11.39	0.20 (0.33)		11.48	0.20 (0.70)	
	ARO-18	11.32	0.31 (0.40)		11.42	0.31 (0.74)	

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3	ARO-8	11.3	0.20 (0.32)		11.41	0.20 (0.70)
4	ARO-16	11.2	0.19 (0.32)		11.29	0.19 (0.69)
5	ARO-7 (O)	12.28	0.20 (0.35)		12.39	0.20 (0.77)
6	ARO-15 (O)	10.99	0.27 (0.36)		11.08	0.27 (0.70)
7						
8	Belalp Cirque (G) (Schindelwig <i>et al.</i> 2012)					
9	Moraine I			12.0±0.6		12.1±0.6
10	VBA-3 (O)	15.08	0.57 (0.66)		15.17	0.57 (1.05)
11	VBA-22	12.49	0.57 (0.63)		12.58	0.57 (0.94)
12	VBA-23	12.57	0.82 (0.87)		12.65	0.82 (1.11)
13	VBA-24	11.95	0.82 (0.86)		12.02	0.82 (1.09)
14	VBA-25	12.04	0.66 (0.71)		12.11	0.66 (0.98)
15	VBA-26	11.09	0.71 (0.75)		11.16	0.71 (0.97)
16	Moraine IV			12.3±0.2		12.4±0.2
17	VBA-6	12.18	0.61 (0.66)		12.25	0.61 (0.96)
18	VBA-15 (O)	9.82	0.94 (0.97)		9.88	0.94 (1.11)
19	VBA-16	12.43	0.65 (0.70)		12.51	0.65 (0.99)
20	Moraine V			12.2±0.6		12.3±0.6
21	VBA-1	13.36	0.68 (0.74)		13.44	0.68 (1.05)
22	VBA-2	11.77	0.57 (0.63)		11.86	0.57 (0.90)
23	VBA-4 (O)	10.32	0.64 (0.68)		10.38	0.64 (0.88)
24	VBA-5	11.78	0.54 (0.60)		11.85	0.54 (0.88)
25	VBA-11	12.16	0.66 (0.72)		12.23	0.66 (0.99)
26	VBA-12	12.78	0.71 (0.77)		12.85	0.71 (1.05)
27	VBA-13	11.85	0.64 (0.70)		11.92	0.64 (0.95)
28	VBA-14	12.02	0.61 (0.66)		12.1	0.61 (0.95)
29	Moraine VI			10.7±0.1		10.8±0.1
30	VBA-17	10.52	0.47 (0.53)		10.58	0.47 (0.78)
31	VBA-18	10.53	0.55 (0.60)		10.59	0.55 (0.83)
32	VBA-19	10.74	0.51 (0.57)		10.81	0.51 (0.82)
33	VBA-20	10.96	0.61 (0.66)		11.03	0.61 (0.90)
34						
35	Great Aletsch (Schindelwig <i>et al.</i> 2012)					
36				12.6±0.8		12.7±0.8
37	VBA-7	13.14	0.74 (0.80)		13.24	0.74 (1.08)
38	VBA-8	13.36	0.54 (0.62)		13.46	0.54 (0.97)
39	VBA-9	11.83	0.97 (1.00)		11.93	0.97 (1.20)
40	VBA-10	12.07	0.74 (0.79)		12.14	0.74 (1.04)
41						
42	GREAT ALETSCHE (Kelly <i>et al.</i> , 2004)					
43				12.3±0.2		12.4±0.2
44	AG-1	12.45	0.51 (0.59)		12.54	0.51 (0.91)
45	AG-2	11.69	0.81 (0.84)		11.78	0.81 (1.07)
46	AG-4	13.16	0.48 (0.99)		12.77	0.48 (0.91)
47	AG-5	12.09	0.70 (0.82)		13.49	0.70 (1.07)
48						
49	Steingletscher (H) (Schimmelpfennig <i>et al.</i> 2014)					
50	Outer moraine			11.0±0.2		11.1±0.2
51	STEI-27	11.18	0.20 (0.32)		11.26	0.20 (0.69)
52	STEI-11	10.91	0.19 (0.31)		10.99	0.19 (0.67)
53	Inner moraine			10.6±0.2		10.7±0.2
54	STEI-8	10.81	0.26 (0.35)		10.90	0.26 (0.69)
55	STEI-9	10.65	0.29 (0.37)		10.73	0.29 (0.69)
56	STEI-10	10.36	0.23 (0.33)		10.44	0.23 (0.66)
57	Lower moraine			10.1±0.5		10.2±0.5
58	STEI-19	10.43	0.15 (0.28)		10.51	0.15 (0.64)
59	STEI-20	9.75	0.08 (0.24)		9.82	0.08 (0.59)
60	STEI-21 (O)	8.72	0.08 (0.22)		8.78	0.08 (0.53)

Val Viola (Hormes et al. 2008)

<i>moraine</i>				12.5 ± 0.5			12.6 ± 0.5	
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<i>VVm4</i>	12.14	0.65 (0.70)			12.23	0.65 (0.70)		
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<i>VVm5</i>	12.8	0.04 (0.30)			12.92	0.04 (0.78)		
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Ferwall Site (Ivy-Ochs et al. 2006)

<i>Schönferwall</i>				13.4 ± 0.4			13.6 ± 0.4	
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<i>F1</i>	13.38	1.01 (1.05)			13.49	1.01 (1.29)		
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<i>F2</i>	14	0.99 (1.04)			14.12	0.99 (1.31)		
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<i>F3</i>	13.02	0.65 (0.71)			13.12	0.65 (1.02)		
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<i>F4b</i>	13.38	0.75 (0.81)			13.47	0.75 (1.10)		
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<i>Kartell</i>				12.1 ± 0.4			12.2 ± 0.4	
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<i>K1</i>	12.32	0.95 (1.00)			12.44	0.96 (1.21)		
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<i>K2</i>	11.58	0.59 (0.64)			11.7	0.59 (0.91)		
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<i>K3a</i>	12.3	1.04 (1.08)			12.42	1.04 (1.28)		
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Kromer Valley (I) (Kerschner et al. 2006; Moran et al. 2016b)

<i>Verhulf-Litzer moraine system</i>				10.6 ± 0.4			10.6 ± 0.4	-60
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<i>KR-1</i>	10.67	0.49 (0.54)			10.75	0.49 (0.80)		
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<i>KR-2</i>	11.12	0.49 (0.56)			11.2	0.50 (0.83)		
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<i>KR-203</i>	10.68	1.27 (1.28)			10.76	1.27 (1.42)		
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<i>KR-303</i>	9.86	0.41 (0.46)			9.94	0.41 (0.71)		
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<i>KR-205 (O)</i>	9.32	0.47 (0.51)			9.39	0.47 (0.73)		
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<i>KR-305 (O)</i>	8.27	0.35 (0.40)			8.34	0.35 (0.60)		
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<i>Scheizer-Kromer moraine system</i>				10.4 ± 0.3			10.5 ± 0.3	-70
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<i>KR-3</i>	10.58	0.50 (0.55)			10.66	0.50 (0.81)		
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<i>KR-4</i>	10.35	0.45 (0.50)			10.43	0.45 (0.76)		
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<i>KR-5</i>	10.82	0.41 (0.48)			10.9	0.41 (0.76)		
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<i>KR-201</i>	10.02	0.92 (0.95)			10.11	0.92 (1.10)		
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Kloster Valley (J) (Moran et al. 2016b)

<i>Kloster valley moraine system</i>				9.8 ± 0.7			9.8 ± 0.7	-70
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<i>KL-1</i>	10.51	0.51 (0.56)			10.58	0.51 (0.81)		
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<i>KL-2A</i>	9.07	0.65 (0.68)			9.14	0.65 (0.85)		
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<i>KL-2B</i>	9.7	0.49 (0.54)			9.78	0.49 (0.76)		
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Falgin Cirque (H) (Moran et al. 2016a)

<i>M1</i>				13.0 ± 0.5			13.1 ± 0.5	
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<i>LT-9</i>	12.96	0.51 (0.59)			13.08	0.51 (0.94)		
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<i>M2</i>				12.1 ± 1.1			12.2 ± 1.1	
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<i>LT-6</i>	11.27	0.43 (0.49)			11.37	0.43 (0.79)		
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<i>LT-7</i>	12.86	0.50 (0.59)			12.99	0.50 (0.93)		
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<i>M3</i>				11.2 ± 0.5			11.3 ± 0.6	-110
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<i>LT-2</i>	12.01	0.46 (0.54)			12.11	0.46 (0.86)		
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<i>LT-3</i>	11.29	0.43 (0.50)			11.38	0.43 (0.80)		
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<i>LT-4</i>	10.89	0.41 (0.48)			10.98	0.41 (0.76)		
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<i>LT-5</i>	10.56	0.40 (0.47)			10.64	0.40 (0.74)		
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La Mare GalcierR (I) (Baroni et al. 2017)

<i>MLm1</i>				12.2 ± 0.4			12.3 ± 0.4	-225
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<i>LAM 13.3</i>	12.21	0.41 (0.50)			12.32	0.41 (0.85)		
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<i>MLm2</i>				10.7 ± 0.2			10.8 ± 0.2	-225
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<i>LAM 14.1</i>	10.79	0.83 (0.86)			10.87	0.83 (1.05)		
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<i>LAM 14.2</i>	10.56	0.52 (0.57)			10.64	0.52 (0.81)		
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<i>MLm3</i>				10.9 ± 0.3			11.0 ± 0.3	-140
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<i>LAM 13.1</i>	10.63	0.63 (0.67)			10.72	0.63 (0.89)		
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<i>LAM 13.2</i>	11.12	0.51 (0.56)			11.2	0.51 (0.83)		
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MLm4				10.5±0.6			10.6±0.6	-140
	LAM 13.4	11.17	0.51 (0.57)		11.26	0.51 (0.84)		
	LAM 13.5	10.35	0.48 (0.54)		10.43	0.48 (0.78)		
	LAM 13.6	9.92	0.40 (0.45)		10	0.40 (0.71)		
Careser Glacier (J) (Baroni <i>et al.</i> 2017)								
MC2				10.1±0.4			10.2±0.4	-180
	POVE 3	10.08	0.42 (0.48)		10.16	0.42 (0.74)		
MCI				10.1±0.9			10.1±0.9	-160
	POVE 1	10.71	0.42 (0.48)		10.73	0.42 (0.76)		
	POVE 2	9.42	0.43 (0.48)		9.51	0.43 (0.71)		

For Review Only