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Latest Pleistocene and Holocene glacier variations in the European Alps

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ABSTRACT

In the Alps, climatic conditions reflected in glacier and rock glacier activity in the earliest Holocene show a strong affinity to conditions in the latest Pleistocene (Younger Dryas). Glacier advances in the Alps related to Younger Dryas cooling led to the deposition of Egesen stadial moraines. Egesen stadial moraines can be divided into three or in some cases even more phases (sub-stadials). Moraines of the earliest and most extended advance, the Egesen maximum, stabilized at 12.2 ± 1.0 ka based on ^{10}Be exposure dating at the Schönferwall (Tyrol, Austria) and the Julier Pass-outer moraine (Switzerland). Final stabilization of moraines at the end of the Egesen stadial was at 11.3 ± 0.9 ka as shown by ^{10}Be data from four sites across the Alps. From west to east the sites are Piano del Praiet (northwestern Italy), Grosser Aletschgletscher (central Switzerland), Julier Pass-inner moraine (eastern Switzerland), and Val Viola (northeastern Italy). There is excellent agreement of the ^{10}Be ages from the four sites. In the earliest Holocene, glaciers in the northernmost mountain ranges advanced at around 10.8 ± 1.1 ka as shown by ^{10}Be data from the Kartell site (northern Tyrol, Austria). In more sheltered, drier regions rock glacier activity dominated as shown, for example, at Julier Pass and Larstig valley (Tyrol, Austria). New ^{10}Be dates presented here for two rock glaciers in Larstig valley indicate final stabilization no later than 10.5 ± 0.8 ka. Based on this data, we conclude the earliest Holocene (between 11.6 and about 10.5 ka) was still strongly affected by the cold climatic conditions of the Younger Dryas and the Preboreal oscillation, with the intervening warming phase having had the effect of rapid downwasting of Egesen glaciers. At or slightly before 10.5 ka rapid shrinkage of glaciers to a size smaller than their late 20th century size reflects markedly warmer and possibly also drier climate. Between about 10.5 ka and 3.3 ka conditions in the Alps were not conducive to significant glacier expansion except possibly during rare brief intervals. Past tree-line data from Kaunertal (Tyrol, Austria) in concert with radiocarbon and dendrochronologically dated wood fragments found recently in the glacier forefields in both the Swiss and Austrian Alps points to long periods during the Holocene when glaciers were smaller than they were during the late 20th century. Equilibrium line altitudes (ELA) were about 200 m higher than they are today and about 300 m higher in comparison to Little Ice Age (LIA) ELAs. The Larstig rock glacier site we dated with ^{10}Be is the type area for a postulated mid-Holocene cold period called the Larstig oscillation (presumed age about 7.0 ka). Our data point to final stabilization of those rock glaciers in the earliest Holocene and not in the middle Holocene. The combined data indicate there was no time window in the middle Holocene long enough for rock glaciers of the size and at the elevation of the Larstig site to have formed. During the short infrequent cold oscillations between 10.5 and 3.3 ka small glaciers (less than several km^2) may have advanced to close to their LIA dimensions. Overall, the cold periods were just too short for large glaciers to advance. After 3.3 ka, climate conditions became generally colder and warm periods were brief and less frequent. Large glaciers (for example Grosser Aletschgletscher) advanced markedly at 3.0–2.6 ka, around 600 AD and during the LIA. Glaciers in the Alps attained their LIA maximum extents in the 14th, 17th, and 19th centuries, with most reaching their greatest LIA extent in the final 1850/1860 AD advance.

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1. Introduction

Downwasting of the large piedmont glaciers back into the Alpine valleys during the Alpine Lateglacial was interrupted by a series of successively smaller glacier advances, which terminated around the end of the Younger Dryas cold phase (summary in Ivy-Ochs et al., 2006b, 2008). These sets of glacier advances (stadials) were already recognized a century ago by Penck and Brückner (1901/09) and since then have been essentially refined and redefined in numerous regional studies. A variety of methods have been used to assign single moraines or entire moraine systems to one of these stadials or sub-stadials. This set of methods includes relative topographic and stratigraphic position of the moraines, moraine morphology (shape, freshness, boulder size), glaciological characteristics of the paleoglaciers, ELA (equilibrium line altitude) depression relative to the LIA (Little Ice Age), ELA in the catchments and a comparison with the ELA of similar glaciers in the vicinity. More recently, isotopic dating with cosmogenic radionuclides of boulders on moraines in Switzerland, Austria, Italy and France has shown that the assignment of moraines to the particular stadials with traditional methods seems to be largely correct and meaningful (Ivy-Ochs et al., 1996, 2006a,b; Kelly et al., 2004; Federici et al., 2008; Hormes et al., 2008).

During most of the Holocene, glaciers in the Alps were never significantly larger than they were during the LIA. Therefore, the LIA extent stands as a suitable reference dimension for paleoglaciological comparisons. The lateral moraines that were last occupied by the LIA glaciers – most of them during the 1850 AD advance – are frequently composed of stacked till units with a complex stratigraphy (Schneebeli and Röthlisberger, 1976; Röthlisberger and Schneebeli, 1979; Röthlisberger et al., 1980; Röthlisberger, 1986; Furrer et al., 1987; Furrer, 1991; Furrer and Holzhauser, 1989). Detailed morphological mapping of the moraines, soil stratigraphy in lateral and terminal moraines, lithostratigraphy of peat bogs close to glacier tongues, pollen analysis, dendrochronology, and lichenometry were combined with archaeological and historical methods to track changes in glacier size during the Holocene. A combination of moraine stratigraphy, the lithostratigraphic study of peat bogs in close vicinity to the glacier tongues, with inwash of inorganic sediment associated with periods of glacier high stands, and pollen analysis proved to be one of the most successful combinations of techniques for reconstructing a sequence of past glacier advances (Patzelt and Bortenschlager, 1973). The assignment of absolute ages in the Holocene is largely based on radiocarbon dating and, more recently, also on dendrochronology (Nicolussi and Patzelt, 2000).

Climatic warming since the mid 1980s led to rapid shrinkage of glacier tongues. Since then, wood fragments and peat have been exposed in situ or washed out from subglacial sedimentary basins at a number of glaciers. Radiocarbon and dendrochronological dating of this material indicates that the glaciers were much smaller and tree line was higher than today in the early and middle Holocene (Slupetzky, 1988, 1993; Slupetzky et al., 1998; Nicolussi and Patzelt, 2000, 2001; Hormes et al., 2001; Nicolussi et al., 2005; Joerin et al., 2006, 2008).

In the context of Lateglacial and Holocene climate change research, rock glaciers (creeping mountain permafrost; Haeberli, 1985) also play an important role. They are phenomena of discontinuous alpine permafrost and as such good indicators for the mean annual air temperature (MAAT) for the period they are active (Barsch, 1996). At the lower boundary of discontinuous alpine permafrost, the MAAT is approximately $-2\,^{\circ}$C. In the Alps, there are broadly three altitudinal belts of rock glacier distribution (Haeberli, 1983; Frauenfelder and Kääb, 2000). Presently active rock glaciers, rock glaciers that have become inactive during the past few decades/centuries, as well as a few smaller relict rock glaciers are found in the uppermost altitudinal belt. The intermediate and the lower rock glacier belts are comprised of relict rock glaciers. Rock glaciers in the lowermost belt are for the most part morphologically associated with moraines of the Younger Dryas glacier advance (Kerschner, 1978; Sailer and Kerschner, 1999; Frauenfelder and Kääb, 2000; Frauenfelder et al., 2001). The age of the intermediate altitude rock glaciers, which are found about 200–300 m lower than the upper belt, is difficult to constrain. Reliable radiocarbon dating is impossible due to the lack of suitable organic material, and the rock glaciers are clearly too old for lichenometric dating. Based on soil stratigraphy at a location in the Larstig valley (Tyrol, Fig. 1), Heuberger (1966) suggested a mid-Holocene age for rock glaciers of the middle altitudinal belt there. This would have required a significant drop in temperatures, thus a marked mid-Holocene cold pulse, for at least several centuries at around 7.0 ka for rock glaciers to be active at 2200 m a.s.l. in Larstig valley. To verify or contradict the postulated mid-Holocene age of the Larstig rock glacier phase, we dated boulders at the type locality with cosmogenic ^{10}Be.

In this paper we present a summary of the evidence for suggested periods of glacier advance during the final phase of the Alpine Lateglacial and the Holocene. We interweave data obtained from ^{10}Be surface exposure dating, radiocarbon dating of wood and peat washed out from the presently melting glacier tongues, dendrochronological investigations on wood from the glacierized basins, tree-line studies and archaeological evidence. In such a framework, the new ^{10}Be exposure dates for the Larstigtal rock glacier site are critical, as they provide clear evidence against a prolonged mid-Holocene cold oscillation (estimated age 7.0 ka) in the Alps. Details on Holocene glacier fluctuations can be found in Nicolussi and Patzelt (2000, 2001), Hormes et al. (2001, 2006), Joerin et al. (2006, 2008) and Holzhauser et al. (2005). Timberline fluctuations are comprehensively treated in Nicolussi et al. (2005). Ivy-Ochs et al. (2006a,b, 2008) provide an overview of Lateglacial glacier fluctuations.

2. Geographic setting

2.1. The Alps

The European Alps are situated close to the southern fringe of the European mainland. They rise close to the Mediterranean seaboard in southern France, trending in a northerly direction towards the Mont Blanc massif. From there they trend in an easterly direction across Switzerland and Austria, finally ending at the river Danube in the outskirts of Vienna (Austria). In the Tirol region (Austria and Italy), the Alps reach their largest N–S extent of roughly 300 km. The highest peaks are well above 4000 m a.s.l. These are situated in western Switzerland and neighbouring southeastern France (highest peak of the Alps: Mt. Blanc, 4809 m a.s.l.). According to the World Glacier Inventory (http://nsidc.org/data/docs/noaa/g01130_glacier_inventory/), at this time there are 5345 glaciers in the Alps, found principally in the higher elevation regions of Switzerland, Italy, Austria, and France. Most Alpine glaciers are small, 81% are smaller than 0.5 km^2 and only 1% are larger than 8.4 km^2 (Zemp, 2006). Presently, the westerlies are an important source of moisture for the Alps especially in the northern regions. Low-pressure systems in the Mediterranean are important for prolonged spells of cold and wet conditions, particularly along the southern fringe of the Alps. Annual precipitation along the northern and southern fringes are rather similar, but due to the topographic barriers in the north and south, the inner Alpine valleys are usually rather dry (Fliri, 1974, 1984; Frei and Schär, 1998). Air temperatures are presently rather mild and usually similar to

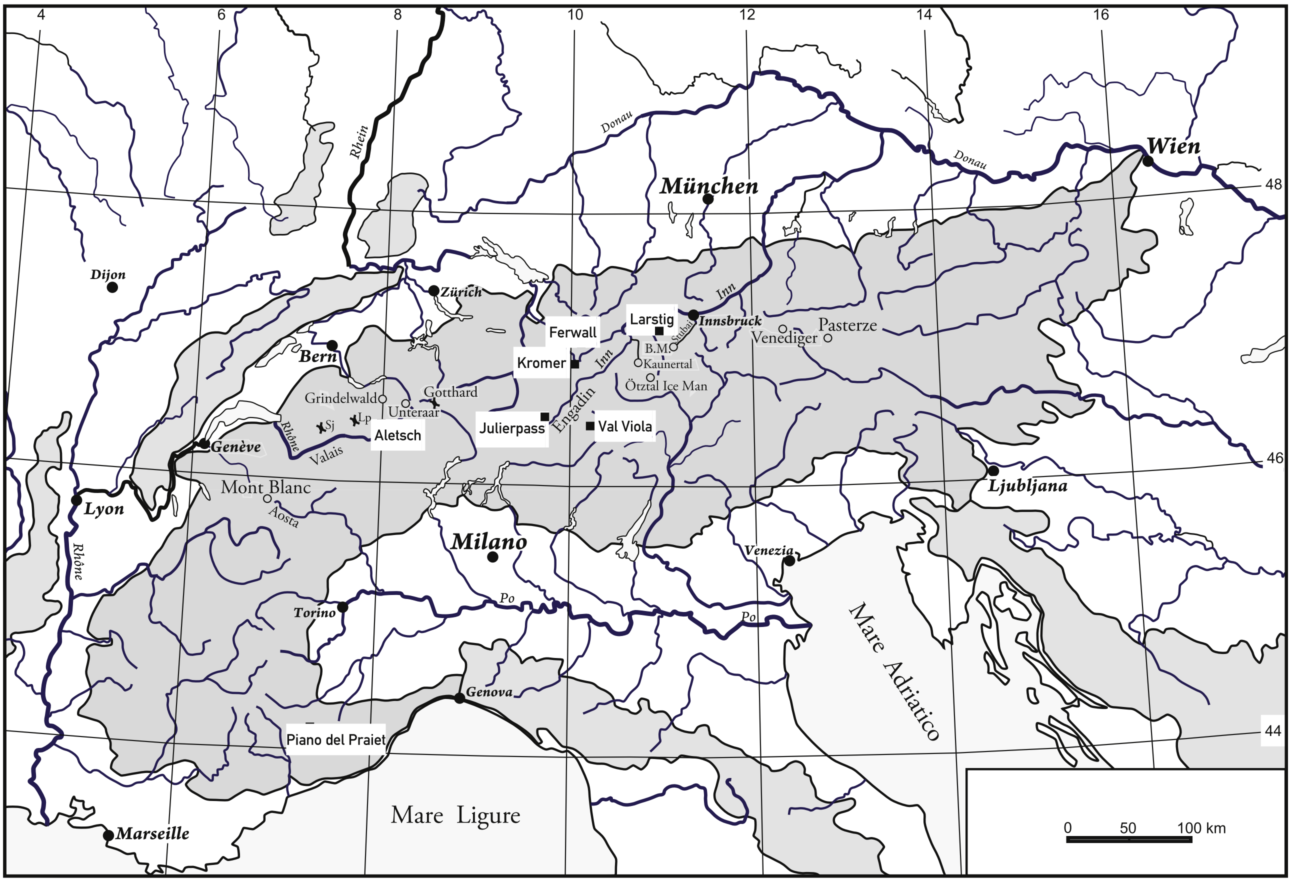

Fig. 1. Reference map for site locations discussed in the text. Areas above 700 m a.s.l. shown as grey shaded pattern. Sj indicates Schnidejoch, Lp Lötschenpass, B.M. Buntes Moor.

those from sites outside the Alps at similar altitudes. Summer temperatures at the ELA of the Alpine glaciers are usually positive, and hence most of the Alpine glaciers are temperate glaciers.

2.2. Larstigtal

Larstigtal is a small, northward trending tributary valley of the Ötz valley in the northwestern Stubai Mountains in Tirol (Austria), centered at 47°8′ N and 11°1′ E, about 30 km southwest of Innsbruck (Fig. 1). The valley is situated entirely in crystalline bedrock of the Stubai-Ötztal massif composed of granites, amphibolites and gneisses of the Ötztal-Stubai complex (Hammer, 1929). The timberline (*Pinus cembra*, *Larix decidua*), which marks approximately the 7 °C summer isotherm (Körner, 2007), is at 2150–2200 m a.s.l. Annual precipitation in Larstig valley is about 1000 mm/year (Tirol-Atlas, 2007) with a clear summer maximum.

In the head of the valley, a small north-facing glacier (Larstigferner) still exists (Fig. 2). According to the Austrian Glacier Inventory from 1969, the steady-state ELA was then at 2840 m a.s.l. (Gross, 1983). During most of the recent summers, the annual EL was higher than the highest point of Larstigferner, which is at 3080 m a.s.l. During the LIA, the eastern tongue of Larstigferner advanced on top of a still active rock glacier. At that time, the ELA of Larstigferner dropped to 2720 m a.s.l., which is rather typical for similarly oriented glaciers in the surrounding area. Relict rock glacier deposits are found along the western side of the valley where they originate from the east-facing scree slopes. These deposits were first described by Heuberger (1954), then by Senarclens-Grancy (1958) and finally, with a high degree of detail, by Heuberger (1966). Whereas in the older papers the deposits were described as moraines, Heuberger (1966) was the first to interpret them as rock glaciers.

We studied the two lowermost rock glaciers in the valley. They are 250–350 m long and 100 m wide along their central flow lines. Both exhibit well developed concentric transverse ridges and furrows (Fig. 3) typical for rock glaciers (e.g. Martin and Whalley, 1987; Barsch, 1996; Humlum, 1998). Surface morphology shows that their general trend of movement was towards the longitudinal axis of the valley and then downvalley. The surface slope is about 20°. The frontal slopes of the rock glaciers are about 10–20 m high and inclined at ca. 30°, which is approximately the angle of repose of the material. The upper margins of the rock glaciers at the foot of the bare rock faces are draped by talus. The lowermost rock glacier is covered by flat slabs of gneiss, which are 0.5–1 m in diameter. Farther up, larger boulders (1–4 m) dominate. The relatively flat frontal angle, the concave-upward profile (Fig. 3), and the fact that the upper slopes are draped by scree indicate that the rock glaciers are relict (Barsch, 1996). Based on the length of the rock glaciers and assuming an average creep velocity of 0.5 m a^{-1} (Barsch, 1977; Hoelzle et al., 1998; Kääb et al., 2003, 2007; Haeberli et al., 2006) with lower velocities at the inception of movement, we estimate that the rock glaciers took 500–1500 years to reach their present size.

In Larstigtal the maximum extent of the Egesen stadial (Younger Dryas) is delineated by lateral and end moraines with a glacier terminus at 2000 m a.s.l. (Fig. 2). The ELA of the glacier was at about 2480 m a.s.l., 240 m lower than the LIA ELA. At the point where the left lateral Egesen moraine is overtopped by the northernmost rock glacier, it is separated from the rock glacier by a relict podsol, which developed on top of the moraine (Heuberger, 1966). A partly eroded podsol is also present on top of the northernmost ridge of the rock

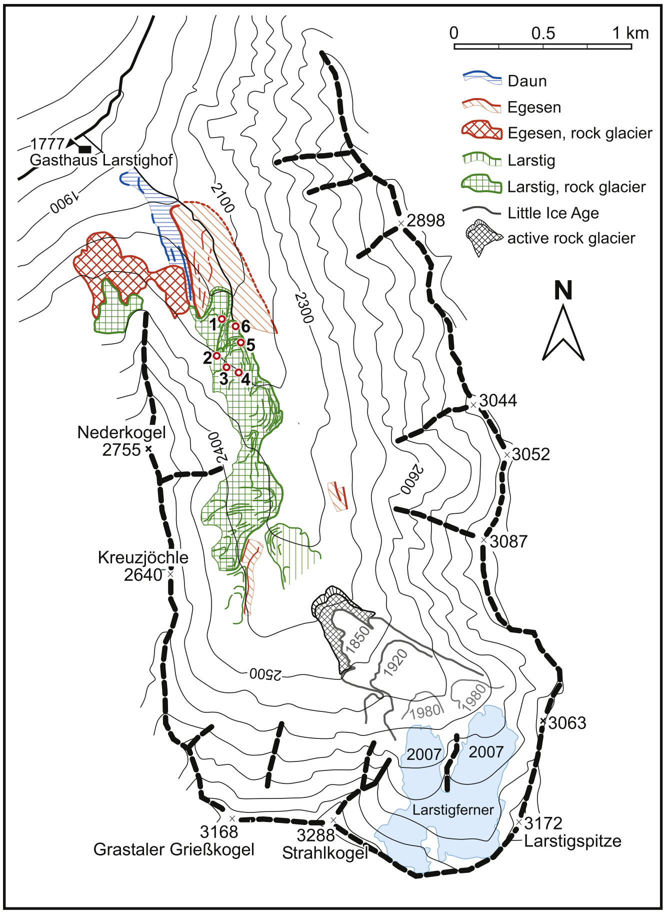

Fig. 2. Geomorphological sketch map of Larstig valley, moraines and rock glaciers after Heuberger (1966). Solid lines indicate moraine and rock glacier ridges, dashed where inferred. Boulders sampled for ^{10}Be surface exposure dating shown with circle and sample number.

glacier. Based on the presence of this paleosol Heuberger (1966) concluded that rock glacier activity, which he named the Larstig oscillation, must have occurred during the postglacial climatic optimum. Later, the period of activity was suggested to be of similar age as the mid-Holocene Frosnitz advance of glaciers in the Venediger Mountains farther to the east (Patzelt and Bortenschlager, 1973).

Fig. 3. Orthophoto draped on laser scan image of Larstigtal (Orthophoto/DEM: © Land Tirol – tiris). Scale bar in lower left corner is 250 m.

3. Methods

We determined surface exposure ages (Gosse and Phillips, 2001) for Larstig rock glacier boulders with cosmogenic ^{10}Be. Six boulders on each rock glacier were sampled (Fig. 2). Rock samples were prepared as described in detail in Ivy-Ochs et al. (2006b). All accelerator mass spectrometry (AMS) measurements were done at Laboratory of Ion Beam Physics at ETH Zurich (Synal et al., 1997). Surface exposure ages (Table 1) were calculated using a sea level, high latitude ^{10}Be production rate of 5.1 ± 0.3 atoms per gram SiO_2 per year, with a muon contribution of 2.2% (Stone, 2000). Scaling to the sample latitude (geographic) and altitude follows Stone (2000). No correction for past geomagnetic field variations was necessary (Masarik et al., 2001). Ages are corrected for an erosion (rock surface weathering) rate of 3 mm ka^{-1} (as discussed in Ivy-Ochs et al., 2004). Stated errors on individual exposure ages reflect analytical uncertainties only. Four out of the five exposure ages (in years), $10{,}640 \pm 440$ (LAR1), $10{,}240 \pm 410$ (LAR2a), $10{,}400 \pm 710$ (LAR3), $10{,}680 \pm 820$ (LAR4), cluster well and give a mean of 10.5 ± 0.8 ka. The age of LAR5 (9080 ± 680 years) is somewhat younger, implying it may have experienced post-depositional toppling or spalling. LAR6 was not measured successfully.

Original radiocarbon dates are found in the references given. Uncalibrated radiocarbon ages are given as ^{14}C yr BP. Radiocarbon dates have been calibrated at the 2 sigma level using OxCal 4.0 (Bronk Ramsey, 2001) with the IntCal04 data set (Reimer et al., 2004), they are given as ka cal age ranges.

Depressions of the ELA are calculated relative to the LIA ELA in the respective catchments (Gross et al., 1977). Steady-state ELAs are calculated from glacier maps with an accumulation area ratio (AAR) of 0.67. Unless stated otherwise, "modern" or "present-day" climatic conditions refer to the period 1961–85 AD. At that time, many glaciers in the Alps reached an equilibrium condition and readvanced until the early 1980s (Patzelt, 1986; cf. Joerin et al., 2008). Gletscher, Kees, Ferner, and Vadret are all words for glacier in the local language (the first three German and the last Rumantsch).

4. Latest Pleistocene to Little Ice Age glacier variations

4.1. 12.9–10.5 ka

Following the Bølling/Allerød interstadial, glaciers in the Alps readvanced markedly during the Younger Dryas cold period (Patzelt, 1972; Maisch, 1981). Glacier advances during the Egesen stadial (Heuberger, 1966, 1968; Maisch, 1981, 1982; see references in Kerschner et al., 2000) were traditionally linked to the Younger Dryas, which was later verified by exposure dating at the Julier Pass site (Ivy-Ochs et al., 1996). The marked climate instability of the 1300 year-long Younger Dryas (Alley, 2000) is reflected by the fact that two and in some cases even three distinct groups of moraines were constructed (Maisch, 1981, 1987). ^{10}Be surface exposure dates have been published from six Egesen moraines (Table 2, Fig. 4). The maximum extent was reached before 12.3 ± 1.6 ka (Julier Pass, outer moraine) and 12.2 ± 1.0 ka (Schönferwall) when these moraines stabilized (Ivy-Ochs et al., 2006b). The two oldest ages of the Piano del Praiet moraine in the Maritime Alps (Federici et al., 2008) are also in the 12 ka range, although the mean age for the moraine is 11.3 ± 0.9 ka (Federici et al., 2008). Moraines of the second phase of the Egesen stadial, which includes Julier Pass (inner moraines, 11.3 ± 0.9 ka), Val Viola (Alp Dosdé moraine, 11.2 ± 1.0 ka; Hormes et al., 2008), and possibly also moraine of the Grosser Aletschgletscher (11.2 ± 1.0 ka; Kelly et al., 2004; Federici et al., 2008), stabilized around the end of the Younger Dryas.

Table 1
Boulder information, AMS-measured ^{10}Be concentrations and calculated exposure ages for the rock glacier boulders at Larstigtal, Austria (47.13°N, 11.01°E).

Boulder No.	Alt. (m a.s.l.)	Thickness (cm)	Shielding corr.[a]	^{10}Be 10^4 atoms/gram[b]	Exposure age (years)[c]	Exposure age, Erosion-corrected[d] (years)
LAR1	2130	3	0.937	26.82 ± 1.10	10,360 ± 420	10,640 ± 440
LAR2a	2190	4	0.921	26.26 ± 1.05	9980 ± 400	10,240 ± 410
LAR3	2200	3	0.938	27.54 ± 1.88	10,130 ± 690	10,400 ± 710
LAR4	2200	2	0.938	28.50 ± 2.21	10,390 ± 800	10,680 ± 820
LAR5	2175	2	0.938	23.94 ± 1.79	8880 ± 670	9080 ± 680

[a] Shielding correction includes both the dip of the sampled surface and the shielding due to the surrounding topography (following Dunne et al., 1999). We assume an exponential drop off of production with depth into the rock with an attenuation length of 157 g cm^{-2} and a rock density of 2.65 g cm^{-3}.
[b] AMS measurement errors are at the 1σ level, including the statistical (counting) error and the error due to the normalization to the standards and blanks. Sample ratios are normalized to the in-house standard S555 with a nominal ratio of 95.5×10^{-12} based on a ^{10}Be half-life of 1.51 Ma.
[c] Age errors include analytical uncertainties only.
[d] Exposure ages have been corrected for a boulder surface erosion (weathering) rate of 3 mm ka^{-1}.

A moraine set found between the innermost Egesen stadial moraines and the LIA moraines was the basis for naming of the Kromer stadial by Gross et al. (1977), because these moraines were particularly well developed in the Kromer valley (Fig. 1). The exposure age of the Kromer moraine is discussed below. Similar moraines in the neighbouring Ferwall group have been used for the naming of the Kartell stadial (Fraedrich, 1975; Sailer, 2001). They are located several hundred meters outside the LIA moraines but several kilometres upvalley from the Egesen maximum terminal position. A Preboreal age was estimated for those moraines (Gross et al., 1977). The ELA depression of the Kromer moraines is in the order of 80 m, while that of the Kartell moraines is 120 m. Moraines at similar positions were also reported from the Engadin (Beeler, 1977; Maisch, 1981; Suter, 1981). At the Kartell site (Fig. 5), ^{10}Be surface exposure ages from three boulders give a mean age for moraine stabilization of 10.8 ± 1.0 ka (Ivy-Ochs et al., 2006b), which supports the Preboreal age. Whether all the moraines found at a position intermediate between the LIA and inner Egesen moraines were deposited during the early Preboreal, as was the Kartell moraine, remains an open question.

In several pollen records a climatic reversal, known as the Palü oscillation (Alp Palü, Engadin, Switzerland), was recognized by an increase in non-arboreal pollen during the earliest Holocene (Burga, 1987; Zoller et al., 1998). Based on a radiocarbon date of 9460 ± 140 BP, it was inferred that the Palü glacier advanced beyond the LIA 1850 AD position just before 11.18–10.30 ka cal. A post-Younger Dryas cold oscillation in a similar time range was identified in paleobotanical records from several sites in both the Swiss and Austrian Alps (Wick and Tinner, 1997; Burga and Perret, 1998; Haas et al., 1998; Tinner and Kaltenrieder, 2005). In a peat bog at timberline near Alpeiner Ferner (Stubai Mountains, Tyrol; Weirich and Bortenschlager, 1980), no cold reversal was found after the start of organic sedimentation. The basal date at the Alpeiner Ferner site is 9630 ± 95 ^{14}C yr BP (11.22–10.71 ka cal). Although the ages are in a quite similar range, the Palü oscillation must be older based on biostratigraphical reasons. A similar proposed Preboreal cold phase, the Piottino oscillation (Zoller et al., 1966) turned out to be equivalent to the Younger Dryas at the type locality (Küttel, 1977). Exact radiocarbon dating in this time window (around 9600 ^{14}C yr BP) is hampered by the existence of the 500-year-long radiocarbon plateau (Becker and Kromer, 1986, 1993). Nevertheless, the discussed cold oscillation(s) reflect the unstable climate conditions of the earliest Holocene.

At the end of the Younger Dryas and in earliest Holocene increasingly dry conditions led to glacier downwasting. Yet conditions remained relatively cold and rock glacier systems developed. According to the new ^{10}Be data presented here (Table 1), the Larstig rock glaciers were active around 10.5 ± 0.8 ka (see Section 5. Discussion below).

4.2. 10.5–3.3 ka

Around 10.5 ka there was a distinct shift towards generally warmer and likely also drier conditions that lasted almost uninterruptedly until about 3.3 ka. Glaciers were smaller than today for most of the time and forest growth was possible in areas that are presently ice covered. Glaciers advanced during a few short phases, the magnitude of the advance depending on the size of the glacier. Small glaciers are thin and steep with relatively simple topography and have a relatively short response time (years to decades) to

Table 2
Average ^{10}Be exposure ages for Egesen stadial moraines and younger sites in the European Alps (sites shown in Fig. 1).

Site	Country	Latitude	Approx. elevation (m a.s.l.)	ΔELA (m)	Mean ^{10}Be exposure age (years)	Number of boulders	Reference
Egesen							
Schönferwall	A	47.06	1730	−290	12,200 ± 800 (1000)[a,b]	4	Ivy-Ochs et al., 2006b
Julier Pass (outer)	CH	46.47	2220	−250, Suter, 1981	12,300 ± 1400 (1600)[a]	4	Ivy-Ochs et al., 1996, 2006b
Julier Pass (inner)	CH	46.47	2200	−250, Suter, 1981	11,300 ± 600 (900)[a]	3	Ivy-Ochs et al., 1996, 2006b
Val Viola	I	46.41	2140	−220, Burga, 1987	11,200 ± 800 (1000) [a,b]	2	Hormes et al., 2008
Grosser Aletsch	CH	46.39	2065–2135		11,200 ± 800 (1000)[a,b]	4	Kelly et al., 2004
Piano del Praiet	I	44.13	1810	−260 to 320	11,300 ± 400 (900)	4	Federici et al., 2008
Early Holocene							
Kartell	A	47.07	2200	−120	10,800 ± 800 (1000)[a,b]	3	Ivy-Ochs et al., 2006b
Julier rock glacier	CH	46.47	2160		10,400 ± 400 (700)[a]	1	Ivy-Ochs et al., 2006b
Larstig rock glacier	A	47.13	2200		10,500 ± 600 (800)[a]	4	this work
Kromer	A	46.91	2180	−75	8400 ± 400 (700)	5	Kerschner et al., 2006

ΔELA refers to the ELA depression for the paleoglacier with respect to the LIA ELA. Mean landform ages are averages, not error-weighted means. The uncertainty of the average moraine age is based on the probability density distribution of all ages with their errors for that site. The uncertainty in parenthesis includes the error of the ^{10}Be production rate.
[a] Corrected for erosion with a rate (3 mm ka^{-1}) (as in original references).
[b] Corrected for 30 cm of snow (density 0.3 g cm^{-3}) coverage for 4 months.

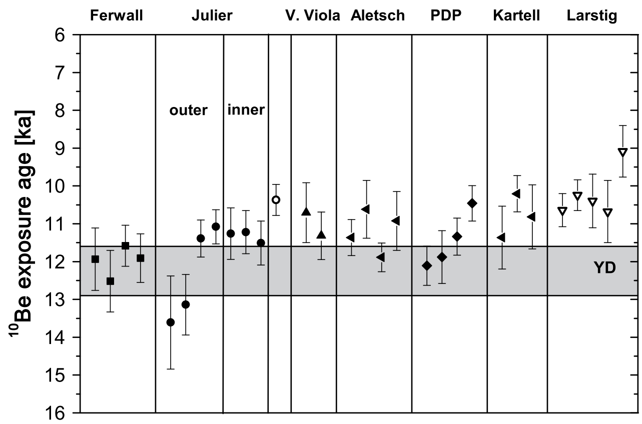

Fig. 4. Plot of all published ^{10}Be exposure ages for Egesen and early Holocene sites in the Alps. Filled symbols are dates from boulders on moraines, open symbols are dates from boulders on rock glaciers. PDP stands for Piano del Praiet. YD is Younger Dryas. Average ages and references for each site are given in Table 2.

climatic perturbations (Jóhannesson et al., 1989a,b). In contrast, large glaciers need several decades to a century or more to build up the ice mass for a significant advance. During the period 10.5–3.3 ka some glaciers with sizes of a few km^2 or less advanced briefly to approximately their subsequent LIA extents. Larger glaciers however remained well behind their LIA extents. In general, glaciers fluctuated on a scale delimited by extents of 1850 and extents smaller than those of today (Nicolussi and Patzelt, 2000, 2001). This period of generally glacier-hostile climatic conditions is documented at several of the larger glaciers in the Alps (Porter and Orombelli, 1985; Burga, 1991, 1993; Baroni and Orombelli, 1996; Orombelli and Mason, 1997; Nicolussi and Patzelt, 2000, 2001; Hormes et al., 2001, 2006; Joerin et al., 2006, 2008).

Radiocarbon data indicate that varve formation in Silvaplanersee (Engadin, Switzerland) ceased after 9415 ± 90 ^{14}C yr BP (11.09–10.31 ka cal). This points to a marked decrease of glacier activity or even absence of glaciers in the upper catchment of Val Fex until 3.3 ka when varves were deposited again in the lake (Leemann and Niessen, 1994; cf. Ariztegui et al., 1996; Ohlendorf, 1998). This requires an ELA rise of at least 100–200 m relative to the modern ELA (Maisch et al., 1999). Right around this time the Julier

Fig. 5. View of the exposure-dated moraine at Kartell (location shown as Ferwall in Fig. 1). Moraine extends from left center edge to the lower center right of the picture and curves back upvalley. ^{10}Be exposure age 10.8 ± 1.1 ka (Table 2). Photo: S. Ivy-Ochs in 2008.

Pass and Larstig rock glaciers finally stabilized. Similarly, by 10.2 ka or perhaps already by 10.5 ka (Fig. 6) Pasterze (location in Fig. 1), the largest glacier in the Austrian Alps, was smaller than it is today and trees were growing in the area presently still covered by ice (Nicolussi and Patzelt, 2001).

During most of the period between 10.5 and 3.3 ka, timberline was higher than today. Nicolussi et al. (2005) identified several periods when tree line in Kaunertal (Fig. 1) was 120–165 m higher than the 1980s level (Fig. 6). Between 9.04 and 8.52, 7.99–7.80, 7.67–7.57, 7.45–6.32, 5.46–5.30 and 4.74–4.54 ka cal, tree line was higher than the tree species line of 2000 (Nicolussi et al., 2005). Similar results were obtained from the Swiss Alps (Tinner et al., 1996; Tinner and Ammann, 2001; Tinner and Theurillat, 2003; Wick and Tinner, 1997; Joerin et al., 2008). Based on radiocarbon and dendrochronological dating of wood washed out of the Vadret da Tschierva (Engadin, Fig. 1), Joerin et al. (2008) identified three Holocene warm intervals; one around 9.2 ka cal, another from 7.45 to 6.65 ka cal and a third from 6.20 to 5.65 ka cal (Fig. 6). They estimated that during those periods the ELA of Vadret da Tschierva must have been at least 220 m higher than it is today and 320 m higher than it was during the LIA. Those periods with high ELAs correspond well to periods when tree line was up to 165 m above the modern tree line in Kaunertal (Nicolussi et al., 2005). Joerin et al. (2006) report 12 phases of marked glacier recession in the Swiss Alps during the Holocene (Fig. 6), which are inferred from radiocarbon dates on wood and peat fragments washed out from the presently downwasting glacier tongues. The total duration of these phases is estimated as 5100 years. If the Pasterze data (Nicolussi and Patzelt, 2000) are added, the total number of years when glaciers were smaller than today during the Holocene goes up to 5700–5800.

The generally glacier-hostile climate was interrupted by a few, comparatively short, phases of glacier-friendly conditions. The Venediger oscillation (Patzelt, 1972; Patzelt and Bortenschlager, 1973) was dated to be older than 9.2 ka (Patzelt in Nicolussi and Patzelt, 2001), which would be just prior to the first Holocene warm interval (Fig. 6) of Joerin et al. (2006, 2008). Similarly, a moraine of the Glacier de Tsidjore Nouve at Arolla (Valais, Switzerland; Röthlisberger, 1976) was radiocarbon dated to 8400 ± 200 ^{14}C yr BP (10.11–8.78 ka cal). The data from Joerin et al. (2006) suggest that this cold phase was confined to the period between 9.6 and 9.3 ka cal. But at that time, Pasterze glacier was smaller than it is today (Nicolussi and Patzelt, 2001).

A two-phased advance period centered around 8.85 and 8.40 ka cal has been suggested. At that time, smaller glaciers reached approximately their LIA extent, but larger glaciers barely exceeded their present-day extent (Nicolussi and Patzelt, 2001). According to the ^{10}Be ages (mean: 8.4 ± 0.7 ka), the Kromer moraines in the Silvretta Mountains stabilized about this time (Kerschner et al., 2006). However, the glacier in Kromertal was clearly much larger than it was during the LIA, which contradicts the evidence from other sites in the Alps. In the Venediger Mountains, a small glacier advanced to its LIA position shortly after 7.7 ka (Patzelt in Nicolussi and Patzelt, 2001), but tree line remained high. This time period (8.4–7.3 ka) corresponds to the multi-phased Misox oscillation which was identified from pollen data at sites in Ticino (Switzerland; Zoller, 1958, 1960). Patzelt and Bortenschlager (1973) described the Frosnitz oscillation (Venediger Mountains, 7.2–6.8 ka), which was tentatively correlated with the Larstig cold phase, based on the radiocarbon date of 6130 ± 130 ^{14}C yr BP from buried wood at Frosnitzkees. However, based on the date (7.31–6.68 ka cal) the Frosnitz oscillation falls right into the second Holocene warm interval (7.45–6.65 ka cal) of Joerin et al. (2008).

There is evidence for advances of smaller glaciers during the two-phased Rotmoos oscillation (Fig. 6) between about 6.3 and

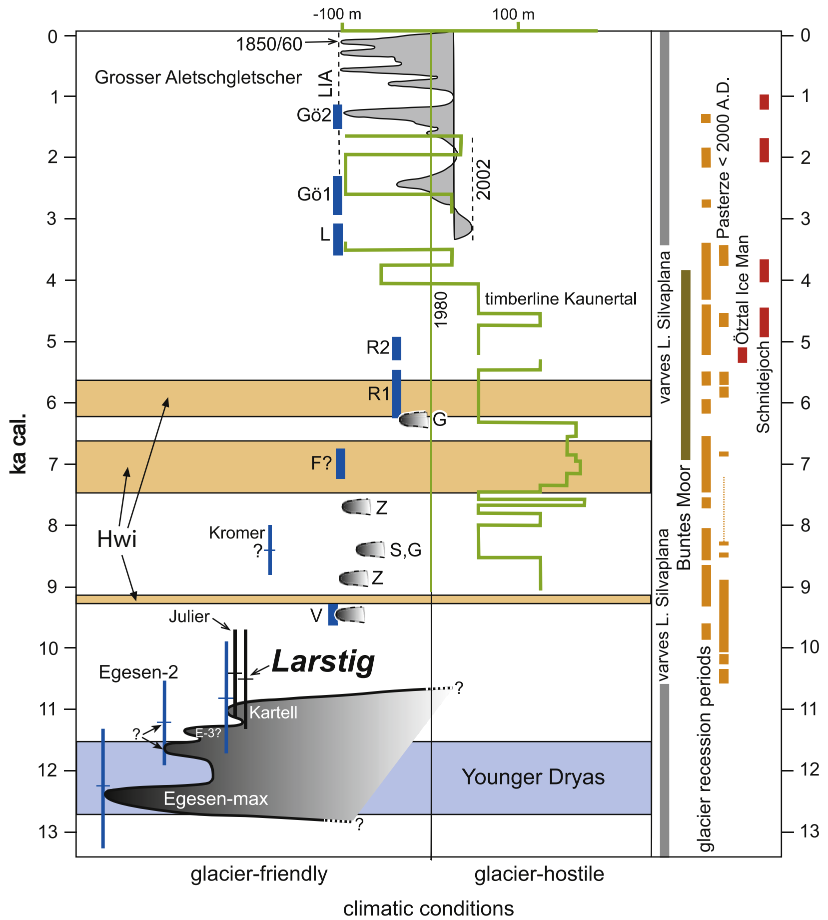

Fig. 6. Summary of glacier variations and general climate evolution during the Younger Dryas and the Holocene. On the left-hand side of the figure: Interpreted time–distance glacier extent for Egesen and early Holocene glaciers from Maisch et al. (1999). ^{10}Be surface exposure ages for Egesen-max (12.2 ± 1.0 ka), Egesen-2 (11.3 ± 0.9 ka) and the Kartell (10.8 ± 1.0 ka) moraines are shown as blue crosses. Julier (10.4 ± 0.7 ka) and Larstig (10.5 ± 0.8 ka) rock glacier ^{10}Be stabilization ages are shown as black crosses (Table 2). Reported glacier advances to the LIA extents or less for the following glaciers are shown as dashed lines. **G:** Gepatschferner (Kaunertal); **S:** Simonykees and **Z:** Zettalunitzkees (Venediger Mountains, Nicolussi and Patzelt, 2001). Short blue bars indicate inferred cold oscillations **V**: Venediger; **F?**: Frosnitz; **R1**, **R2**: Rotmoos 1 and 2; **L**: Löbben oscillation; **Gö1, Gö2**: Göschener 1 and 2 oscillations (see text for references). Hwi: Holocene warm intervals (orange bands) based on data from Vadret da Tschierva (Engadin; Joerin et al., 2008). Green line indicates variations in timberline in Kaunertal (Nicolussi et al., 2005). Advances of Grosser Aletschgletscher from 3.5 ka to present are taken from Holzhauser et al. (2005). On the right-hand side of the figure: Varves present in Silvaplanersee sediment (Engadin; Leemann and Niessen, 1994), as indicator of active glaciers in the catchment area. Buntes Moor, uninterrupted organic sedimentation (Nicolussi and Patzelt, 2001). Glacier recession periods (central and eastern Swiss Alps) from Joerin et al. (2006). Pasterze glacier smaller than at 2000 (Nicolussi and Patzelt, 2000, 2001). Ötztal Ice Man, glacier at Tisenjoch as small as in 1991 (Bonani et al., 1992, 1994; Baroni and Orombelli, 1996). Radiocarbon-dated archaeological materials found at Schnidejoch (Grosjean et al., 2007).

5.0 ka (Bortenschlager, 1970, 1984; Patzelt and Bortenschlager, 1973; Patzelt, 1977), but larger glaciers still remained small. This period roughly corresponds to the Piora oscillation in the Swiss Alps (Zoller, 1960, 1977; Zoller et al., 1966; Heitz, 1975; Bircher, 1982; Renner, 1982) which was identified from pollen data. In any case, the suggested glacier advances were only short interruptions during a period of generally small ice extent (Nicolussi and Patzelt, 2001; Joerin et al., 2006).

At around 5.30–5.10 ka cal (Bonani et al., 1992, 1994), no ice existed at Tisenjoch (3200 m a.s.l.), where the Ötztal Ice Man was found melting out of glacier ice in 1991. At Schnidejoch (2730 m a.s.l.), a pass between the Swiss cantons of Bern and Valais, archaeological evidence dates to 4.90–4.45 ka cal and 4.06–3.70 ka cal (Suter et al., 2005a,b). During these times this direct north–south route was not blocked by the Tungelgletscher (Grosjean et al., 2007). Similar dates (4.30–3.71 ka cal) were obtained from radiocarbon dating of three bows found at Lötschenpass (2690 m a.s.l., Fig. 1) in 1933 (Bellwald, 1992). They indicate that Lötschenpass was also a passable north–south route during that period, and was not blocked by the Lötschengletscher.

After 3.8 ka, smaller glaciers advanced to their LIA extent, but larger glaciers still remained small (Patzelt and Bortenschlager, 1973; Gamper and Suter, 1982; Holzhauser, 1995; Hormes et al., 1998, 2001, 2006; Holzhauser et al., 2005; Joerin et al., 2006, 2008). At Buntes Moor (Tyrol, Fig. 1), the deposition of silty layers, indicating glacier high stands in the immediate vicinity, begins around 3.8 ka after a phase of uninterrupted organic sedimentation since 7.0 ka (Weirich and Bortenschlager, 1980; Nicolussi and Patzelt, 2001). This marks the onset of the Löbben oscillation (3.8–3.4 ka) (Patzelt and Bortenschlager, 1973; Patzelt, 1977). Advances during this time period, based on radiocarbon-dated wood entrained in moraines or buried paleosols, have been reported from numerous glaciers across the Alps (Patzelt and Bortenschlager, 1973; King, 1974; Schneebeli and Röthlisberger, 1976; Zoller, 1977; Röthlisberger et al., 1980; Bircher, 1982; Bless, 1982; Gamper and Oberhansli, 1982; Gamper and Suter, 1982; Renner, 1982). It is likely that at this time the lowermost sediment of the tens of meters high lateral 'LIA' moraines at many alpine glaciers was deposited (Schneebeli and Röthlisberger, 1976; Small, 1983). Many of them were formed during numerous glacier advances with periods of tree growth, soil formation and erosion in between. An example is the right lateral moraine at Steingletscher shown in Fig. 7. Radiocarbon dating suggests this moraine may have begun to accumulate already at about 3 ka (King, 1974).

4.3. 3.3 ka to Little Ice Age

After 3.3 ka timberline moved down to lower altitudes, prolonged glacier advances became more frequent and periods of recession were shorter, leading finally to the LIA advances from the 14th century AD until 1850/60 (Patzelt, 1973; Zumbühl, 1980; Gamper and Suter, 1982; Grove, 1988; Zumbühl and Holzhauser, 1988; Holzhauser et al., 2005). The record of variations of the Grosser Aletschgletscher, which begins at 3.5 ka (Holzhauser et al., 2005) exemplifies the overall trends in the Alps (Fig. 6). The Grosser Aletschgletscher, Unterer Grindelwaldgletscher and Gornergletscher (Valais, Switzerland) advanced between 3.0 and 2.6 ka, culminating at 2.6 ka (Holzhauser, 1995, 1997; Holzhauser et al., 2005; Röthlisberger and Oeschger, 1961). Similar glacier advances during the Göschener I oscillation (3.0–2.3 ka) (Zoller et al., 1966) were recorded in many regions of the Alps (Patzelt and Bortenschlager, 1973; Schneebeli and Röthlisberger, 1976; Röthlisberger et al., 1980; Suter, 1981; Bircher, 1982; Bless, 1982; Renner, 1982; Furrer and Holzhauser, 1989; Deline and Orombelli, 2005). Increased glacier activity in the catchment of Silvaplanersee is reflected in renewed varve formation after 3.3 ka (Leemann and Niessen, 1994).

Numerous Roman artefacts found at Schnidejoch point to an ice-free period between 2.11 and 1.74 ka cal (Suter et al., 2005a,b). Data from the Grosser Aletschgletscher describe a continuous mild interval during Roman times until about 200 AD (Holzhauser, 1997; Holzhauser et al., 2005). Glaciers were estimated to have been smaller around this time than they were at 1930/1920 AD (Nicolussi and Patzelt, 2001; Schlüchter and Joerin, 2004).

In the late Roman times and the early Middle Ages (Göschener II oscillation: Zoller, 1960; Zoller et al., 1966) numerous glaciers in the Alps advanced (Schneebeli and Röthlisberger, 1976; Röthlisberger et al., 1980; Bircher, 1982; Bless, 1982; Orombelli and Porter, 1982; Renner, 1982; Furrer and Holzhauser, 1989; Orombelli and Mason, 1997; Deline and Orombelli, 2005). The Grosser Aletschgletscher, Unterer Grindelwaldgletscher and Gornergletscher advanced at around 500–600 AD and possibly around 800–900 AD (Holzhauser et al., 2005). During the first phase of Göschener II, glaciers in the Eastern Alps may have reached sizes roughly comparable to their final LIA maximum extents (Patzelt, 1977; Nicolussi and Patzelt, 2001).

Nicolussi and Patzelt (2001) describe the conditions of the 9th and 10th centuries as unfavourable for glaciers (cf. Grove and Switsur, 1994). This is corroborated by the data from Schnidejoch. There, the youngest artefacts date to 1230 ± 50 ^{14}C yr BP (670–930 AD) and 1195 ± 50 ^{14}C yr BP (690–970 AD) (Suter et al., 2005b) indicating an ice-free pass (Grosjean et al., 2007). Minor advances around 1100–1200 AD could be distinguished at the Grosser Aletschgletscher and Gornergletscher (Holzhauser, 1995, 1997; Holzhauser et al., 2005). Historical records indicate expansion of the Unterer Grindlewaldgletscher during the 12th century (Zumbühl, 1980; Zumbühl and Holzhauser, 1988; Holzhauser and Zumbühl, 1996).

Taking the Grosser Aletschgletscher and Gornergletschers as representative, maximum LIA ice extents in the Swiss Alps were reached in the 14th, 17th and 19th centuries (Holzhauser et al., 2005; cf. Nussbaumer et al., 2007). The Unterer Grindelwaldgletscher showed similar behaviour (Steiner et al., 2008), but its furthest extent was attained at 1600–1640 AD (Zumbühl, 1980; Zumbühl and Holzhauser, 1988; Holzhauser and Zumbühl, 1999; Holzhauser et al., 2005).

Glacier reaction patterns were similar in the Austrian sector (Patzelt and Bortenschlager, 1973; Patzelt, 1977). Both Pasterze and Gepatschferner advanced in the middle 15th century, the early 17th century and at 1852/1856 AD (Nicolussi and Patzelt, 2000). Maximum extents were reached in the 1850/1860 AD phase. Most Italian glaciers and also many glaciers in the Western Alps reached their LIA maximum extent around 1820 AD and readvanced to almost this extent in 1850 AD (Kinzl, 1932; Orombelli and Porter, 1982; Porter, 1986; Orombelli and Mason, 1997), while glaciers farther to the east were smaller in 1820 AD than in 1850 AD.

Fig. 7. View of right lateral 'Little Ice Age' moraine of Steingletscher located near Gotthard in Fig. 1. Scale is given by the person standing at the lake shore (circled). The complex stratigraphy visible especially in the upper parts of the lateral moraine hints at the period required for the deposit to accumulate, which may have began already by about 3.0 ka (King, 1974). The hill in the left foreground just above the boulder is the end moraine of the most recent advance measured from 1970 until 1989 (http://www.geo.unizh.ch/wgms/). The deposit in the right center is an ice-cored, inset lateral moraine left by the continuous retreat of nearly 300 m since 1990. The active glacier is out of the picture to the right. Photo: M. Maisch in 2008.

5. Discussion

Egesen stadial moraines were built as glaciers expanded during the Younger Dryas cold period. The moraines that formed during advances in the early part of the Younger Dryas (Egesen maximum) mark the furthest extent of ice. These moraines stabilized around 12.2 ± 1.0 ka based on ^{10}Be exposure dates from the Schönferwall and Julier Pass sites (outer moraine). Final stabilization of the inner moraines occurred sometime before 11.3 ± 0.9 ka. This is the average of the ^{10}Be results from Julier Pass (inner moraines, Ivy-Ochs et al., 1996, 1999), Val Viola (Hormes et al., 2008), and the Piano del Praiet (Federici et al., 2008) Egesen moraines (Table 2). Here we have also included the data of the Grosser Aletschgletscher Egesen lateral moraine, although it is classified as an Egesen maximum moraine (Kelly et al., 2004). We have included the Aletsch data in the younger set because at the location where the four boulders were sampled the lateral moraine constructed during the maxium advance may have been reoccupied during a later Egesen stadial advance. Within the given error bands the dates from all six Egesen moraines clearly overlap the Younger Dryas chronozone (Fig. 4). Nevertheless considering the relationship between the ages of the moraines and the general development of Younger Dryas climate in central Europe (e.g. Ammann et al., 2000; Schaub et al., 2008), the set of ages seems to be slightly too young. It may be that several centuries passed before the moraines finally stabilized after glacier downwasting. Another possibility is that we have underestimated the role of snow cover, especially at Aletsch where several of the sampled boulders are located in a lee position on the left lateral moraine where significant snow might accumulate. By and large, winter snow cover, in particular on top of the large boulders, is thin due to wind action and the generally low winter precipitation. In addition, the period with snow cover lasts only a few months. Nevertheless, the depth of snow and periods of snow cover during the Holocene are difficult to assess.

Tentative palaeoclimatic modeling shows that humid conditions prevailed along the northern slope of the Alps in western Austria and eastern Switzerland during the first phase of the Younger Dryas (Kerschner and Ivy-Ochs, 2008). In the interior valleys and in the south-central part of the Alps climate was drier than today, while close to the Mediterranean seaboard more humid conditions prevailed (Federici et al., 2008). After the Egesen maximum advance and in the final phases, climate became successively drier. Glaciers were starved and large rock glacier systems developed in areas of the Alps where peaks remained close to or below the ELA, but cirque floors were above the lower limit of discontinuous permafrost (Kerschner, 1982; Sailer and Kerschner, 1999; Frauenfelder and Kääb, 2000; Frauenfelder et al., 2001). Some of the rock glaciers appear to be multi-phased (Kerschner, 1982; Sailer, 2001), but in detail it is usually impossible to delineate clear interruptions in their kinematic history. Where local topography permits, rock glaciers extended down to 1700–1900 m a.s.l. (Kerschner, 1983, 1985, 1993), indicating the widespread presence of discontinuous permafrost at that altitude during the Younger Dryas cold phase. As rock glaciers were active at least 500–600 m below the presently active rock glaciers, mean annual temperatures were at least 3.5–4 K lower than today (Kerschner, 1978, 1983, 1985; cf. Haeberli, 1983).

The moraines of the Kartell stadial have an early Holocene exposure age of 10.8 ± 1.1 ka. Although within the given errors these moraines have an age similar to the aforementioned Egesen moraines, the Kartell moraines have been differentiated based on morphostratigraphy (Sailer, 2001). They are not Egesen stadial moraines. Kartell stadial moraines may have formed during a glacier advance related to the Preboreal oscillation (Lotter et al., 1992; Björck et al., 1998: 11.30–11.15 ka; Schwander et al., 2000: 11.4–11.1 ka), which is also represented by a marked drop in tree macrofossils at the altitude of present-day timberline (Tinner and Kaltenrieder, 2005). Based on lake-level fluctuations, Magny et al. (2007) infer that climate during the Preboreal oscillation was more humid for the first time since the end of the Younger Dryas. Assuming annual precipitation similar to today, a summer temperature drop of about 1.5 °C (Magny et al., 2007) would have been sufficient for glaciers to advance to their Kartell stadial terminal positions.

Exposure dates from the Larstig site (10.5 ± 0.8 ka) point to rock glacier activity during the early Preboreal. Because the sediment originates from talus at the base of the rock face and to a minor degree from the Egesen lateral moraine it is possible that boulders could contain inherited nuclide concentrations. Presence of such concentrations would lead to exposure ages that are older than the actual time of landform stabilization (Putkonen and Swanson, 2003; Ivy-Ochs et al., 2007). If the boulders had been accumulating ^{10}Be during their transit in the rock glacier, the ages would get successively older towards the front. But we see no distinct pattern in our data with respect to exposure age and boulder location on the rock glacier. Similarly, it seems unlikely that pre-exposure was acquired in the bedrock wall or as the boulders lay at the foot of the slope (cf. Barrows et al., 2004). Material destined for the rock glaciers accumulated during and after the end of the Lateglacial when the bare rock faces were actively spalling; the debris destined for the rock glaciers was accumulating rapidly. There was little time for inherited nuclide concentrations to build up. In any case, boulders with significant pre-exposure would stand out in the data set (Putkonen and Swanson, 2003; Ivy-Ochs et al., 2007). The ages from four of the five Larstig boulders are indistinguishable within the errors (Table 1).

Field relationships indicate that the rock glaciers in Larstigtal overrode and partially incorporated material of the left lateral Egesen moraine (Fig. 2). This means they moved into an ice-free valley after the Egesen glacier had melted back from its maximum extent. Thus the earliest time for onset of rock glacier activity is around 12–11.6 ka. The size and morphology of the Larstig rock glaciers support one continuous period of activity (Frauenfelder and Kääb, 2000) which, based on an average creep velocity of 0.5 m a^{-1} (Hoelzle et al., 1998; Kääb et al., 2003, 2007), lasted from 500 to 1500 years. This is in agreement with the mean exposure age of 10.5 ka. The simplest explanation is that we have dated the stabilization of the rock glaciers after final melting out of permafrost. Nevertheless, we cannot strictly quantify the time lag between the climate warming itself and the melting of the permafrost in the rock glaciers. Preservation of original surface architecture of the deposit, such as ridges, pits, and furrows, is good evidence that the upper surface remained stable after the permafrost melted out. In addition, as the top surface is composed completely of meter-length blocks with no matrix, the possibility for washing-out of fines and consequent boulder instability is minimal (cf. Ivy-Ochs et al., 2007).

Both the rock glacier just inside the right lateral moraine at Julier Pass (10.4 ± 0.7 ka) and the Larstig rock glaciers (10.5 ± 0.8 ka) stabilized around 10.5 ka. By that time, Pasterze glacier was smaller than today. Stabilization of the rock glacier surfaces may have occurred even a few centuries earlier, perhaps during the improvement of climate right after the Preboreal oscillation. During this rock glacier active phase permafrost existed about 300 m lower than it did during the 20th century. Whether these permafrost occurrences reflect equilibrium conditions and hence a roughly −2 °C MAAT lowering, or a transient disequilibrium during the degradation of Younger Dryas or perhaps even older permafrost, remains an open question.

From 10.5 ka cal until about 3.3 ka cal, high timberline altitudes, high summer temperatures and small glaciers were common. Climatic reversals were infrequent and short. During those events small glaciers were able to advance close to or slightly beyond their LIA limits, while large glaciers with long and flat tongues fluctuated around much smaller extents. The pre-3.3 ka cold phases (summarized in Haas et al., 1998), which had been mainly inferred by pollen analysis, were obviously not strong and/or long enough to cause large glaciers to advance at or even close to their LIA limits. This is interesting with respect to rock glacier development. As rock glaciers are operating on multi-century to multi-millennial timescales, there are practically no time windows between 10.5 ka and 3.3 ka, where rock glaciers of the size of those at Larstig, at altitudes 200–300 m below the presently active rock glaciers, could have developed. This is supported by the ^{10}Be stabilization ages at the Larstig site, which point to a late Younger Dryas/early Holocene period of activity.

Joerin et al. (2008) calculate an ELA rise of 220 m relative to 1960–1985 AD (320 m relative to LIA) at Vadret da Tschierva during the Holocene warm intervals (at 9.2 ka cal, 7.45–6.65 ka cal and 6.20–5.65 ka cal). By comparison, the ELA of Hintereisferner (Tyrol, Austria) rose by 180 m from 2970 m a.s.l. (1961–1985 AD) to 3150 m a.s.l. (1986–2005 AD) (Data courtesy of Institut für Meteorologie und Geophysik, Universität Innsbruck). With the empirical glacier-climate model of Ohmura et al. (1992), Joerin et al. (2008) calculate several scenarios for possible changes in summer temperature and precipitation. With unchanged precipitation, they calculate a summer temperature increase of 1.8 K during the Holocene warm intervals. To obtain a more detailed idea of the climatic conditions we cross-checked the results of Joerin et al. (2008) with the analytical glacier-climate model of Kuhn (1981). The model is based on the heat and mass balance equation and allows the calculation of vertical shifts of the ELA from changes in the terms of the equation (Kuhn, 1989). Here we assume a summer temperature shift of 1.2 K, which is derived from the timberline data from Nicolussi et al. (2005) and a 3 percent increase in global radiation, as can be assumed for the early Holocene (Berger, 1978). We use a vertical temperature lapse rate of $-0.007\ \mathrm{K\,m^{-1}}$ and a vertical accumulation gradient of $1\ \mathrm{mm\,m^{-1}}$ (Kuhn, 1981). With unchanged precipitation, the ELA rise amounts to 190 m, with the assumed increase in global radiation contributing only 20 m to the ELA rise. To achieve an ELA rise of 220 m, accumulation has to be reduced by $150\ \mathrm{mm\,a^{-1}}$, which is very small and clearly within the limits of accumulation from climate data. It is also much less than the interannual precipitation variability, which is in the order of 30 percent (Fliri, 1974). This reduction in accumulation may be interpreted as a somewhat higher frequency of dry winters. The additional rise of about 30 m beyond the 190 m calculated above could as well be achieved by a slight increase of the latent heat flux towards the glacier surface, as warm fair-weather conditions at high altitudes bring about a rise in vapour pressure above 6.11 hPa (saturation at melting ice surfaces; Kerschner, 2000). In total, the results obtained agree very well with those from Joerin et al. (2008).

If we assume increased seasonality during the early and middle Holocene (cf. Davis et al., 2003; Joerin et al., 2008), we may assume in a first approximation that only summer temperatures (June–September) were 1.2 K higher, while there was no change during the rest of the year. Under such an assumption, the mean annual temperature would have been 0.7 K higher than during 1961–1985 AD. Consequently, the lower limit of discontinuous permafrost should have been 100 m higher during the Holocene optima events. Neverthless, many of the presently active rock glaciers (in the highest altitudinal belt) that began activity during the early Holocene (right after downwasting of cirque glaciers) may have still survived the warmest phases.

6. Conclusions

Exposure dates from five widely separated sites, ranging from the Maritime Alps to the Eastern Alps, are remarkably consistent in showing that Egesen stadial glaciers advanced in response to Younger Dryas cooling. Egesen maximum moraines (Schönferwall, Julier Pass-outer moraine) stabilized around 12.2 ka while later Egesen moraines (Piano del Praiet, Grosser Aletschgletscher, Julier Pass-inner moraine, Val Viola) stabilized at about 11.3 ka. Towards the end of the Younger Dryas, climatic conditions became increasingly dry. Glaciers downwasted and rock glacier activity increased. Nevertheless, some glaciers were still larger than their LIA extents in the early Preboreal as shown for example by the results from Kartell where moraines stabilized at 10.8 ka.

About 1000 years after the end of the Younger Dryas, there was a shift towards markedly warmer and probably also drier conditions. These conditions lasted almost uninterruptedly until about 3.3 ka. Large glaciers were already smaller than they are today. Rock glacier activity in all but the uppermost altitudinal belt ceased by approximately 10.5 ka. The ^{10}Be data presented here for the two rock glaciers at Larstigtal show that relict rock glaciers can be quite successfully exposure dated with cosmogenic nuclides. The record of mean annual air temperature inferred from the rock glacier location and consequently information they provide on past climatic conditions can be integrated into existing paleoclimate temporal frameworks.

Between 10.5 and 3.3 ka, glaciers were smaller than they are today and timberline was hundreds of meters higher than the tree species limit of around 2000 AD (Nicolussi et al., 2005). Our estimates of palaeoclimatic conditions during the Holocene warm intervals (at 9.2 ka cal, 7.45–6.65 ka cal and 6.20–5.65 ka cal) support the conclusions made by Joerin et al. (2008). When we infer summer temperature increase based solely on timberline altitude change (Nicolussi et al., 2005), the proposed minimum ELA increase of 220 m during the Holocene warm intervals relative to 1961–85 ADELAs requires – in terms of glacier mass balance – a slight reduction in accumulation. This could most easily be explained by a somewhat higher frequency of dry winters during the Holocene warm intervals. From estimates of past timberline and ELA data it follows that summers were warmer than today and total annual precipitation was slightly less. The likely synoptic-meteorological background for such a climate can be seen in stronger Mediterranean high pressure systems, which expanded further to the north (cf. Schär et al., 2004). Assuming as well increased temperature contrasts between summer and winter, the mean annual temperature may have been only about 0.7 K higher than today during the Holocene warm intervals, and the lower limit of discontinuous permafrost about 100 m higher. This is an additional argument against a mid-Holocene phase of low-altitude rock glacier formation.

Between 10.5 and 3.3 ka there were brief, infrequent climate reversals. At those times smaller glaciers advanced to roughly their LIA limits, but larger glaciers remained well behind their LIA extents. Even during the few periods of inferred glacier advances, timberline remained high. A possible explanation for this could be a greater importance of increased accumulation as a driver for the necessary glacier positive mass balance, as has been observed during recent decades in south-western Norway (Nesje and Dahl, 2003; Lie et al., 2006; Nesje, 2006). In any case, there was no time window long enough to permit the development of rock glaciers the size of those at Larstig, which are 200–300 m below the presently active rock glaciers in the Alps. This is consistent with the Preboreal age for the rock glaciers of the intermediate altitudinal belt as determined here. The rock glaciers in Larstigtal stabilized no later than 10.5 ka. They did not form during a mid-Holocene cold oscillation.

Around 3.3 ka timberline moved to lower altitudes, prolonged glacier advances became more frequent and periods of recession were shorter, leading finally to the LIA advances from the 14th century AD until 1850/60 AD. This change towards a more glacier-friendly climate marks a shift to climatic conditions similar to those observed in the Alps during the past few centuries.

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