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Surface exposure dating of moraines in the Kromer valley (Silvretta Mountains, Austria) – evidence for glacial response to the 8.2 ka event in the Eastern Alps?

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Abstract: ^{10}Be dating of a pronounced glacier advance ('Kromer Stadial') in the western part of the Silvretta Mountains (western Austria) yielded ages ranging from 8010 ± 360 to 8690 ± 410 years with a mean age for moraine stabilization of 8410 ± 690 years. Hence, the Kromer Stadial, which was previously considered as early Preboreal, may represent the glacier response to the early phase of the '8.2-ka event', as it is recorded in the Greenland ice-cores. The glacier advanced to a position beyond its 'Little Ice Age' (LIA) limit; the end moraines are situated 750–1000 m downvalley from the glacier ends in 1850. The corresponding drop of the equilibrium line altitude relative to the LIA datum was 75 m. The glacier advance was followed by a period of glacier recession and rock glacier development. In total, the climatic fluctuation may have lasted for about 500 ± 200 years. The age and duration of the climatic fluctuation is in good accordance with the 'Misox cold phase' and CE-3 climatic fluctuation as recorded in Switzerland. Climate during the Kromer Stadial was characterized by more humid conditions than today along the northern fringe of the Alps and slightly negative to moderately positive changes in precipitation in the central part of the Austrian Alps. Summer temperatures changes were most likely in the order of -1 to 0K. Mean annual temperature was 1.5K to 2K lower than today, at least during the second phase of the climatic fluctuation.

Key words: Surface exposure dating, cosmogenic isotope, beryllium-10, 8.2-ka event, glacier–climate models, Austrian Alps, Holocene.

Introduction

The cold phase around 8200 years, the '8.2-ka' or '8k' cold event (Alley *et al.*, 1997; Alley and Ágústsdóttir, 2005 and references therein), is commonly regarded as the most severe climatic downturn in the North Atlantic region during the early Holocene. It was embedded in a longer period of general climatic decline, which began at about 8600 years and ended after 8000 years (Rohling and Pälike, 2005). The event was probably caused by a rapid drainage of proglacial Lakes Agassiz and Ojibway through Hudson Bay (Teller *et al.*, 2002; Clarke *et al.*, 2004) towards the North Atlantic. The outbreak probably started around 8470 cal. BP (Barber *et al.*, 1999) with a dating uncertainty of more than a century to many centuries. According to the Greenland ice core records (Johnsen *et al.*, 2001), the cold phase reached its peak around 8200 years. The ice-core records show a two-pronged nature of the cold event, this is also as expected from climate modelling

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under reasonable assumptions (Renssen *et al.*, 2001). The event left its footprint at various sites in the Northern Hemisphere (Mayewski *et al.*, 2004 and references therein) with the most significant climatic impact being found on the European side of the North Atlantic (Klitgaard-Kristensen *et al.*, 1998; Risebrobakken *et al.*, 2003; Nesje *et al.*, 2004).

In the circumalpine region, numerous lines of evidence point towards a phase of cooler and wetter climatic conditions interrupting a thermal optimum. This includes $\delta^{18}O$-data from lake sediments (von Grafenstein *et al.*, 1998), expansion of less drought-resistant vegetation in central Europe (Haas *et al.*, 1998; Tinner and Lotter, 2001), changes in the chironomid assemblages in alpine lakes (Heiri *et al.*, 2003), the pollen-analytically inferred Misox cold phase (Zoller, 1977; Zoller *et al.*, 1998; Wick and Tinner, 1997) and lake-level highstands with intermediate phases of low lake levels in the Jura Mountains, the Swiss Plateau and the French Préalpes (Magny *et al.*, 2001, 2003a,b; Magny and Bégéot, 2004). In Scandinavia and Eastern Europe, changes in lake sediments and treeline variations (Karlèn, 1976; Barnett *et al.*, 2001; Seppä and Birks, 2002; Veski *et al.*, 2004) show a pronounced cooling. Glaciers in southern Norway advanced during the 'Finse event' to roughly the order of magnitude found during the 'Little Ice Age' (LIA) (Nesje and Dahl, 1991; Dahl and Nesje, 1996; Matthews *et al.*, 2000; Nesje *et al.*, 2000; Nesje *et al.*, 2004), but the actual glacier size there seems to depend on the geographical location of the glaciers. A glacier advance in the southern Coast Mountains of British Columbia at that time remained considerably smaller than the LIA extent (Menounos *et al.*, 2004).

To date, there has been no record for a significant glacier advance in the European Alps around that time, even though alpine glacial history shows numerous advances of LIA order of magnitude during the Holocene (Patzelt and Bortenschlager, 1973; Patzelt, 1977; Nicolussi and Patzelt, 2001 and references therein). In contrast, evidence from fossil peat deposits and tree rings indicates that the early Holocene was a period when large Alpine glaciers were almost continuously small; for extended periods they were even considerably smaller than today (Nicolussi and Patzelt, 2000, 2001; Hormes *et al.*, 2001). Only smaller glaciers showed minor early Holocene advances, which remained behind their LIA limits (Nicolussi and Patzelt, 2000, 2001).

Here we present the results of ^{10}Be surface exposure dating of a moraine complex in the western Silvretta Mountains (Vorarlberg, Austria). In the 1970s, it was considered to be early Preboreal in age (Gross *et al.*, 1977), because the consensus was that glaciers had already melted back and readvanced to their LIA extent during the Preboreal (Patzelt and Bortenschlager, 1973). The glacier advance formed well-preserved moraines in many valleys of the western part of the Silvretta Mountains. They occupy an intermediate position between the prominent moraines from the 1850 advance upvalley and the innermost moraines of the Younger Dryas Egesen Stadial (Ivy-Ochs *et al.*, 1996) farther downvalley. The most complete moraine system is located in the Kromer valley; hence the name 'Kromer Stadial' was chosen (Gross, 1974). We also give a palaeoclimatic interpretation of the 'Kromer Stadial' in terms of precipitation change and discuss the possible temperature change associated with the 8.2-ka event in the Austrian Alps. Open towards the west and north, the research area occupies a key position at the humid western fringe of the Austrian central Alps (Figure 1a). Hence, the results of the palaeoclimatic interpretation should be representative for a wider area along the northern and western fringe of the Eastern Alps. The results should be well comparable with those of other studies in the northern foreland of the Alps.

Research area

Kromertal is a short valley running NNE–SSW in the western part of the Silvretta Mountains, Vorarlberg province, Austria (Figure 1). The mouth of the valley is located at 1800 m a.s.l.; the highest peaks are at 3109 m and 3121 m. The neighbouring peaks of the Ferwall Mountains to the north are only slightly higher than the valley floor of the middle part of the valley. Precipitation coming from the W–N sector can therefore easily reach the valley. Consequently, the LIA equilibrium line altitude (ELA) was low in comparison with the more sheltered mountains in the Ötztal or Engadin (Table 1).

The upper part of the valley is divided into two separate cirque areas (Figure 1b). Bedrock is composed of orthogneiss and amphibolites of the Austroalpine Silvretta Nappe (Geologische Karte der Republik Österreich, sheet 169), which tends to form rugged peaks and blocky talus.

Several small glaciers occupy the background of the valley, which formed well-developed moraines during the 1850 (LIA) advance (Figure 1b, Table 1). Within the forefields of the glaciers, moraines of the 1920 advance and the 1980 advance can be observed. Presently, only remnants of the glaciers are left.

Lateral moraines of the Egesen Stadial are present on both sides of the lower part of the valley. During the maximum Egesen advance, a dendritic glacier covered the entire western part of the central Silvretta Mountains with an end position near the village of Parthenen in the Montafon valley (Hertl and Kerschner, 2001; Hertl, 2002). A second substage is documented by lateral moraines. The third Egesen substage ended in an area that is occupied by a reservoir lake (Vermunt-Stausee; just to the north of the area shown in Figure 1b) today. The end moraine of the third substage can be seen on old photographs and is mentioned by Kinzl (1929). In the central part of the Kromer valley, a large lateral moraine exists with the glacier end at 1980 m. It either represents another substage of the Egesen Stadial (Figure 1b,c) or the 'Kartell Stadial' (Fraedrich, 1979; Sailer, 2002). Moraines in similar positions can also be found in other valleys of the western Silvretta Mountains, but are missing in the eastern part of the mountain range (Hertl, 2002).

The moraines of the Kromer Stadial are situated in the surroundings of a small plateau in the middle of the valley at $\pm$2200 m a.s.l., 750–1000 m downvalley from the LIA glacier ends. Two groups of blocky moraines can be distinguished, which are close together. In some places they only consist of blocks up to 5 m long, wedged together and supporting each other. To the northeast of an Austrian Custom's cabin, they rest discordantly on top of the older moraines mentioned above.

In the western part, the frontal moraines of the Kromer maximum are situated at an altitude of 2130 m. The inner moraines (2160 m) are morphologically more prominent and visually dominate the surroundings. On their proximal side, a depression is partly infilled by fine-grained fluvial sediments. In the eastern part of the glacier, the end of the maximum advance is at 2155 m. The right lateral moraine begins at an altitude of 2460 m and can be followed downvalley over a distance of 1100 m until it merges with the end moraine. It is morphologically as fresh as a LIA moraine and is largely composed of angular blocks up to 5 m long. On its upper end it was partly incorporated and overrun by a younger moraine–rock glacier complex, which was slightly larger than LIA.

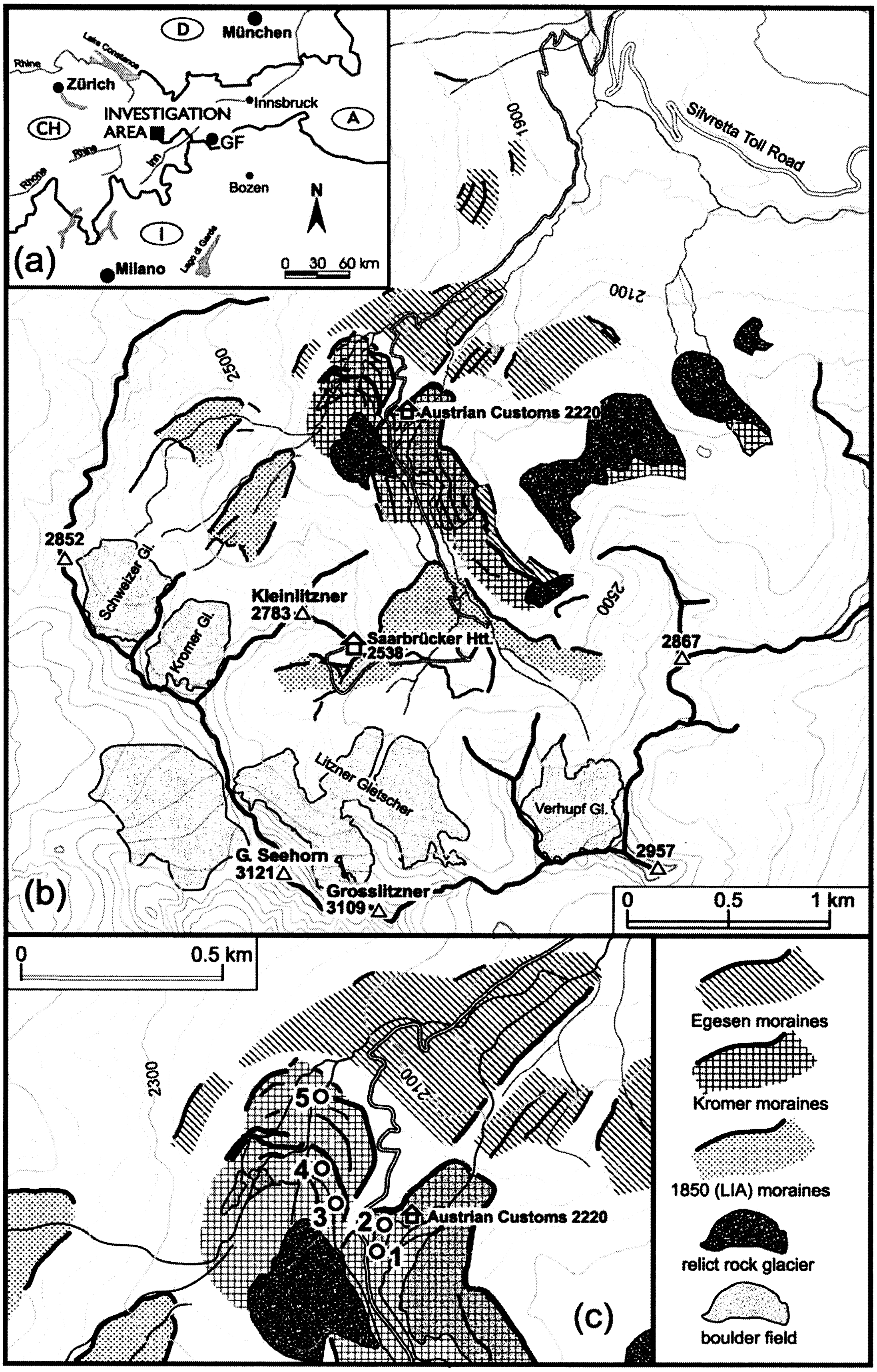

Figure 1 (a) Index map of the investigation area. GF, Gepatschferner. (b) Glaciers, moraines and rock glaciers in the Kromer valley. (c) Location of the sample points. Numbers refer to sample numbers in Table 2. Note: past glacier tongue in front of the Kromer moraines may be younger in age than 'Egesen' (Younger Dryas) (see text)

The area inside the moraines is densely strewn with angular to slightly subangular blocks ('boulder field' in Figure 1), which rest either directly on exposed bedrock or on a thin veneer of till. Rock glacier development can be observed in the moraines and talus cone just northeast of Kleinlitzner down to an altitude of 2200 m. It was more

Table 1 Glaciers in the Kromer valley

Name	ELA 1850 (m a.s.l.)	Surface 1850 (km^2)	Exposure
Schweizer Gletscher	2565	0.474	NE
Kromer Gletscher	2520	0.382	NE
Litzner Gletscher (West)	2630	0.284	NE
Litzner Gletscher (East)	2515	1.029	N
Verhupf Gletscher (West)	2570	0.639	NW

widespread in front of the glaciers in the neighbouring cirques to the east.

Characteristics of the glacier

The equilibrium line altitude of the Kromer glacier was calculated with an accumulation area ratio of 0.67 (Gross *et al.*, 1977) at 2480 m. This is in good agreement with the altitude of the onset of the right-hand lateral moraine at 2460 m. The weighted average LIA-ELA datum level ('Bezugsniveau', Gross *et al.*, 1977) of the catchment was at 2555 m, resulting in an ELA lowering (ΔELA) of 75 m. By comparison, ΔELA of the Younger Dryas (Egesen Stadial) maximum in the western Silvretta Mountains was 380 m and that of the morphologically rather similar Kartell advance in the neighbouring Ferwall Mountains was 120 m (Sailer, 2002).

In total, the glacier covered an area of 4.21 km^2 (1850: 2.81 km^2). The maximum thickness of the eastern part of the glacier was in the order of 100 m, and the 'characteristic thickness' near the former ELA was around 60–80 m. The western part of the glacier was steeper and slightly thinner. Assuming an annual net ablation at the glacier snout, 300 m below the ELA, of 2–3 m water equivalent/yr, the volumetric reaction time (Jóhannesson *et al.*, 1989a,b) of the glacier was in the order of 20–40 years. Even if we consider that the reaction time of the length of a glacier is longer, the glacier was able to react to a climatic perturbation on a timescale of a few decades.

Dating

Based on the 'intermediate' position of the Kromer moraines between the innermost Egesen moraines and the LIA moraines, an early Preboreal age was assumed for the Kromer Stadial (Gross *et al.*, 1977). The latter gives a good idea of the maximum Holocene glacier extent, which had been reached during the Preboreal (Patzelt and Bortenschlager, 1973; Patzelt, 1977) in the Central Alps of Austria.

To date the moraines, we chose surface exposure dating with the cosmogenic radionuclide ^{10}Be (Gosse and Phillips, 2001). We sampled five blocks from different positions on the moraines near the Austrian Custom's cabin (Figure 1c). Samples KR-1 and KR-2 come from angular to slightly subangular blocks in a flat area inside of the left-hand lateral moraine of the eastern glacier tongue. They rest on a thin bed of till with some bedrock outcrops. Samples KR-3 and KR-4 come from two large blocks, which are wedged into a mass of smaller blocks on the proximal side of the end moraine of the second advance of the western glacier tongue. Sample KR-5 comes from a block just inside the end moraine of the first advance of the western tongue. All samples were obtained 1.5–2 m above the surrounding terrain to minimize the influence of snowcover, which usually lasts from November until April/May. An examination of the generally windswept boulder sites shows that snowcover is of minor influence. Fine material is generally absent at the sampling sites, so post-depositional exhumation of the blocks can be excluded.

All samples come from quartz veins in orthogneiss. They only minimally (a few millimetres at best) protrude above the surface of the blocks, which showed no signs of significant postdepositional surface erosion. Sample thickness was around 1–2 cm in all cases; sampling was done with hammer and chisel.

The quartz vein samples were crushed and sieved to a grain size less than 0.6 mm. Quartz was purified and freed of meteoric ^{10}Be using selective chemical dissolution with weak HF (Kohl and Nishiizumi, 1992). After addition of ^{9}Be carrier, the quartz was dissolved with concentrated HF. We used anion and cation exchange and selective pH-controlled precipitations (Ochs and Ivy-Ochs, 1997) to separate and purify Be. ^{10}Be/^{9}Be, along with appropriate standards and blanks, was measured by accelerator mass spectrometry at the ETH/PSI tandem facility in Zurich (Synal *et al.*, 1997). Chemistry blanks were $1-2 \times 10^{-14}$.

To calculate the exposure ages we used a sea-level, high-latitude ^{10}Be production rate of 5.1 ± 0.3 atoms/g SiO_2 per year, with a contribution from muons of 2.6% (Stone, 2000). Scaling to the sample latitude (geographic) and altitude were also done following Stone (2000). No correction has been made for past changes in magnetic field intensity, which would have been minimal at this latitude (Masarik *et al.*, 2001; J. Masarik, personal communication, 2004).

The exposure ages for the five boulders range from 8010 ± 360 to 8690 ± 410 years (Table 2). Boulder ages reflect measurement (analytical) errors only. The average age for the moraine is 8410 ± 690 years. The error is based on the 1σ confidence interval of the mean as well as systematic errors of the production rate (cf. Gosse and Phillips, 2001). Hence, with a probability of 68% the minimum age for moraine stabilization lies within the range of 7720 years to 9100 years. Within the errors, these results show that the moraine may have stabilized around the onset of the 8.2-ka event, approximately contemporaneous with the outbreak of Lake Agassiz at 8470 cal. BP (Barber *et al.*, 1999).

Table 2 Boulder information, AMS-measured ^{10}Be concentration and calculated exposure ages

Boulder no.	Altitude (m a.s.l.)	Height above ground (m)	Shielding corr.[a]	^{10}Be (atoms/g)	AMS measured error (%)[b]	Exposure age (years)
KR-1	2220	1.8	0.983	2.46×10^5	4.8	8410 ± 400
KR-2	2220	1.0	0.966	2.51×10^5	4.7	8690 ± 410
KR-3	2175	1.9	0.961	2.32×10^5	5.0	8380 ± 420
KR-4	2180	1.9	0.982	2.28×10^5	4.5	8010 ± 360
KR-5	2135	1.5	0.958	2.30×10^5	4.0	8550 ± 340

[a]Shielding correction includes both the dip of the sampled surface and the shielding due to the surrounding topography.
[b]AMS measurement errors are at the 1σ level, including the statistical (counting) error and the error resulting from the normalization to the standards and blanks.

Palaeoclimatic interpretation

The aim of this section is to estimate precipitation during the Kromer Stadial under the assumption that the moraines represent a glacier advance, which is contemporaneous with the early phase of the 8.2-ka event. Further, we assume a steady-state glacier. We use the analytical glacier–climate model by Kuhn (1981, 1989; see also Kaser and Osmaston, 2002) and the statistical glacier–climate models by Ohmura *et al.* (1992; 'OKF-model'), Krenke (1975); Khodakov (1975) and the 'Liestøl equation' (Ballantyne, 1989). This comparative approach has already proven useful for the reconstruction of Younger Dryas regional precipitation patterns in the Alps (Kerschner *et al.*, 2000). A more detailed discussion of the models and the necessary climatological calibration can be found in Kerschner (2005). Modern precipitation at the ELA in the Kromer valley, as calculated with the OKF-model, is in the order of 2000–2200 mm/yr, which is slightly more than the values given by Auer *et al.* (2001).

The palaeoclimatic inferences refer to 'modern' climate, here defined as the climate during the middle of the twentieth century (1931–60 WMO normal period), which provides the climatic background for the Austrian glacier inventory. Hence, the ΔELA of 75 m relative to the LIA-ELA has to be corrected for the modern situation. Maisch *et al.* (1999) give an average value of 90 m for the ELA rise in Switzerland and 75 m for the Swiss part of the Silvretta Mountains since the 'Little Ice Age'. Therefore, we use an ΔELA of 150 m relative to an assumed 'modern' ELA as a basis for our calculations.

All models require external information on summer temperature change (ΔT_S). From lake-level fluctuations and related palaeolimnological data, Magny *et al.* (2001, 2003a,b) and Magny and Begeot (2004) infer a shorter vegetation period and a temperature depression of the warmest month of -2K to -2.5K for the 8.2-ka event. We assume that this applies also to summer (June–August) temperature. Heiri *et al.* (2003) infer a ΔT_s of -1K relative to the climate before and after the event from chironomid assemblages in an alpine lake at the northern slope of the Swiss Alps. As summers during the early Holocene were rather warm, the latter value may even be interpreted as no temperature change at all relative to the 1931–60 period. The value given by Heiri *et al.* (2003) is practically the same as a ΔT_s of -0.8K, as reported by Haas *et al.* (1998) for several sites in Switzerland. Von Grafenstein *et al.* (1998) calculate a depression of mean annual temperature of -1.7K from $\delta^{18}O$ measurements in lake sediments in southern Germany close to the northern fringe of the Alps.

Under modern climatic conditions, fields of temperature and temperature change are spatially conservative. Hence, we can rather safely assume that the ΔT_s values, which range between -2.5K and 0K, apply also to the research area.

The results, which are summarized in Table 2 show that all models give practically the same results. The only exception is 'winter precipitation' calculated from the 'Liestøl-equation', which tends to overestimate winter precipitation under present-day alpine conditions (Kerschner, 2005). Precipitation was apparently reduced by about 10–15% for a ΔT_s of -2K, practically similar to modern values or slightly higher than today for a ΔT_s of -1.5K to -1K and in the order of +25% to +35% for a ΔT_s of 0K.

Discussion

The morphology of the moraines and, in particular, the abundance of rock-fall debris, shows that the glacier advance was preceded and accompanied by intense rockfall activity, mainly during the second advance. This may have been caused by increased rockfall on the glacier surface during a short warm phase separating the maximum advance from the younger advances.

The rock glacier-like deformation of the moraines on the northern side of Kleinlitzner down to 2200 m within the former glacier-covered area (Figure 1b, centre) shows that colder conditions prevailed after the recession of the glacier. Similarly, rock glaciers in front of Kromer moraines in the cirques to the east reached down to 2250 m and 2180 m. The lengths of the rock glaciers vary between 250 and 500 m. Assuming a rock glacier deformation rate of 1–2 m/yr (Haeberli, 1985; Barsch, 1996), we can estimate that rock glacier activity lasted for about 200–500 years. As the existence of permafrost was possible down to 2200 m, we can infer a depression of the lower limit of permafrost of about 200–300 m relative to presently active rock glaciers in the vicinity. This is less than the Younger Dryas permafrost depression in the surrounding mountain regions (Kerschner, 1978; Sailer and Kerschner, 1999). On the other hand, it agrees well with postglacial phases of rock glacier development at comparatively low altitudes, as shown by Heuberger (1966). However, they are not necessarily connected to the 8.2-ka event.

Qualitatively, the morphological record shows a first phase of glacier-friendly climatic conditions, which led to a glacier advance with an ELA-depression of 75 m. This was followed by a short phase of glacier recession and increased rockfall activity, and a subsequent period of a multiphased glacier advance with a similar ELA-depression as the first advance. The rock glacier-like deformation of the moraines, which began after the recession of the glacier, indicates prolonged cold and comparatively dry climatic conditions.

According to the exposure dates, the moraine complex was stabilized around 8410 ± 690 years. Considering the errors in the data, the Kromer advance may possibly be correlated with the 8.2-ka event in a broader sense. Whether the advance was a direct response to the 8.2-ka event cannot be said, as this would require a dating accuracy of a few decades or less. If we assume that glacier-friendly climatic conditions and, hence, the advance started shortly after the outbreak of Lake Agassiz around 8470 cal. BP, resulting from an immediate reaction of European climate to the changed conditions in the North Atlantic Ocean, the glacier advance should have lasted for about 100 years or less. At the type locality, this provides sufficient time for the advance, because of the short reaction time of the comparatively small and short glacier involved. How fast a glacier can develop under favourable climatic conditions has recently been shown at Mt St Helens (Schilling *et al.*, 2004). Together with the subsequent phase of rock glacier activity, the climatic downturn should have lasted for several centuries (*c.* 300–700 years), which is in good agreement with the duration of the climatic downturn as given by Haas *et al.* (1998) and Heiri *et al.* (2003).

The age of the moraines was surprising, because the glacier was considerably larger than during the 1850 (LIA) maximum. In the absence of absolute ages, they were therefore considered to be of early Preboreal age (see above). Larger glaciers in the Alps were more or less continuously small during the early Holocene (Nicolussi and Patzelt 2000, 2001, Hormes *et al.*, 2001) and there is ample evidence for glacier extents smaller than today around 6500 BC (8450 years, cf. discussion in Nicolussi and Patzelt, 2001; Hormes *et al.*, 2001). However, Gepatschferner, which is situated in the rather continental Ötztal Mountains, advanced after 6450 BC (8400 years) to a greater extent than in 1950 but clearly less than in the LIA;

at the same time Simonykees in the southern Venediger Group south of the main alpine divide advanced to a position that was larger than in 1870 (G. Patzelt, in Nicolussi and Patzelt, 2001: 69). In total, Nicolussi and Patzelt (2001) document a short two-phased period of minor advances in the drier areas of the Austrian Alps at the Boreal/early Atlantic transition, which peaked around 6900 BC and 6450 BC (8850 and 8400 years). During this period, some glaciers reached advanced positions but remained smaller than around 1850; larger glaciers hardly reached their modern extent (Nicolussi and Patzelt, 2001: 70). The extent of the Kromer glacier was not reached in the central Alps during that period.

If we assume that there is no significant time gap between the termination of the glacier advance and the ^{10}Be ages of the Kromer moraines, we have to try to explain the obvious discrepancy in glacier extent between the oceanic and continental parts of the Eastern Alps.

First, the reaction time of large and long glaciers in the central Alps may have been too long compared with the duration of the climatic downturn. As glaciers were small before the onset of the climatic downturn, a glacier-friendly period of about 100 years may have been too short to build up enough ice for an advance over several kilometres to a position comparable with the Kromer moraines in the western Silvretta Mountains.

Second, the ELA depression in the comparatively dry central Alps could have been less than along the humid northern fringe of the Alps, similar to the situation during the Younger Dryas Egesen Stadial (Kerschner *et al.*, 2000). During the early Younger Dryas, ΔELA in the western Silvretta Mountains was −380 m (Hertl, 2002), whereas it was −210 m in the upper part of Kaunertal valley (western Ötztal Mountains), where Gepatschferner is located (Kerschner, 1979). The difference in ΔELA was therefore 170 m, which can be easily explained by increasingly drier conditions in the central, more sheltered areas (Kerschner *et al.*, 2000).

If we assume from the field evidence (Nicolussi and Patzelt, 2001) that the Kromer Stadial ELA in the central Alps was roughly similar to the LIA-ELA, the difference between the ELA depression at the northwestern fringe (Kromer valley) and in the central part of the Austrian Alps (Kaunertal valley) would have been in the order of 80 m, which is about half of the early Younger Dryas difference in ELA depression. Such a difference in ΔELA is best explained by increased contrasts in precipitation, as fields of temperature depression are spatially conservative.

Under this assumption, precipitation change was calculated for the Gepatschferner region. The results (Table 3) show a drastic decrease in precipitation in the order of −30 to −40% for a ΔT_s of −2K and more. This is more than during the Egesen Stadial (Younger Dryas) maximum advance (Kerschner *et al.*, 2000) and does not seem to be realistic. For ΔT_s in the order of −1K, precipitation change is in the order of −5%. With modern summer temperatures, the necessary precipitation change is in the order of +20%. Taking into account that precipitation at higher altitudes is difficult to measure, we may interpret the results for a moderate temperature depression as 'roughly similar to modern values'.

From this point of view, a summer temperature depression in the order of −2 to −2.5K, as given by Magny *et al.* (2001, 2003a,b) and Magny and Begeot (2004) may be too large for the western part of the Eastern Alps; a ΔT_s in the order of −1K or even no temperature change at all seems to give more realistic results in terms of precipitation change. Such a ΔT_s is also more in accordance with the pollen-analytical record of the Misox cold phase (eg, Zoller, 1977; Wick and Tinner, 1997; Zoller *et al.*, 1998), the CE-3 cold phase by Haas *et al.* (1998) and the cooling as determined by Heiri *et al.* (2003). In addition, there is no evidence for a timberline depression in the order of −300 to −350 m, as it would be caused by a ΔT_s of −2 to −2.5K.

If we accept a ΔT_s in the order of −1 to 0K, the precipitation regime during the Kromer advance was characterized by increased precipitation along the northwestern fringe of the Austrian Alps. In the central areas, precipitation was roughly similar or slightly higher than today. Such a precipitation pattern is typical for a climate dominated by westerly to northwesterly airflow, as suggested by Magny and Bégeot (2004). It would favour lake-level highstands in the western foreland of the Alps (Magny *et al.*, 2003a; Magny and Begeot, 2004), the spread of less drought-resistent vegetation types in central Europe (Tinner and Lotter, 2001) and increased snow cover at higher altitudes (Schmidt *et al.*, 2004).

The spatial distribution of rock glaciers as features of discontinuous permafrost is mainly controlled by the mean annual air temperature (MAAT) (Haeberli, 1974; Barsch, 1996). Hence, the lowering of the MAAT can be inferred from the depression of the rock glacier belt (Kerschner, 1983; Sailer and Kerschner 1999; Frauenfelder *et al.*, 2001). Using a temperature lapse rate of −0.007K/m, the permafrost depression of 200–300 m during the later phase of the Kromer Stadial is equivalent to a ΔMAAT of −1.4K to −2.1K. This

Table 3 Relative precipitation change (% modern) at the altitude of the 'modern' ELA for different scenarios of summer temperature change during the Kromer advance

ΔT_s	Ohmura *et al.* (1992) (%)	Kuhn (1981) (%)	Krenke (1975) (%)	Khodakov (1975) (%)	Liestøl-equation (Ballantyne, 1989) (%)
Kromertal, western Silvretta Mountains, ΔELA = −150 m (field evidence)					
−2.5	−21	−20	−27	−24	−36
−2	−13	−12	−17	−15	−24
−1.5	−4	−3	−7	−6	−11
−1	4	5	5	5	5
−0.5	13	14	18	15	24
0	22	23	31	27	46
Gepatschferner, central Ötztal Mountains, ΔELA = −90 m (assumed)					
−2.5	−41	−49	−42	−37	−43
−2	−30	−34	−31	−27	−33
−1.5	−18	−20	−19	−17	−21
−1	−6	−6	−6	−5	−7
−0.5	7	8	8	7	9
0	19	22	24	20	28

agrees well with the value of -1.7K as given by von Grafenstein *et al.* (1998) for Lake Ammersee in southern Bavaria. The field evidence shows that the relatively strong ΔMAAT occurred at least during the second part of the climatic fluctuation during and after the recession of the glaciers from the terminal moraines.

Conclusions and outlook

A series of closely spaced moraines in the Kromer valley (western Silvretta Mountains, Austria) were deposited by a glacier, which was clearly larger than during the LIA but smaller than the inner substages of the Egesen Stadial (Younger Dryas). The ELA of the glacier was 75 m lower than during the LIA. ^{10}Be dating of five boulders from the moraines gives ages ranging between 8690 and 8130 years, with a mean minimum age for moraine stabilization of 8410 ± 650 years. Hence, the glacier advance may have been caused by the same climatic deterioration, that also caused advances of some glaciers in the central Alps and the 'Finse-event' in southwestern Norway. The proximity of the ages to the start of the 8.2-ka event is striking. However, as the ^{10}Be ages are minimum ages for moraine stabilization, it cannot be completely ruled out that the actual glacier advance occurred during an older phase of glacier-friendly climate in the early Holocene.

Based on the mean age for moraine stabilization and assuming that the glacier advance started shortly after the outbreak of Lakes Agassiz and Ojibway, the glacier advance lasted for about 100 years. The glacier recession from the end moraines was followed by a period of rock glacier development down to altitudes 200–300 m lower than presently active rock glaciers. This indicates that the climatic downturn continued for several centuries after the glacier advance. A total duration for the climatic fluctuation in the order of about 500 ± 200 years seems to be realistic in light of a short glacier fluctuation and a subsequent phase of rock glacier development. This is in good accordance with the Misox/CE-3 cold phase (Haas *et al.*, 1998; Zoller *et al.* 1998).

In the Austrian Alps, a glacier advance far beyond the LIA extent seems to have been limited to areas at the humid northern and western fringe that receive large amounts of precipitation today. In the central, more sheltered areas of the Austrian Alps, glaciers remained somewhat behind their LIA limits. This obvious discrepancy in glacier extent may be explained either by the different reaction times of small, well-nourished glaciers on the one hand and large glaciers with a much longer reaction time on the other. Another possibility is a climatic regime different from that of today, with an increased spatial contrast in precipitation between the central areas and the northern and western fringe of the Alps, caused by a higher frequency of cool westerly to northwesterly airflow. A combination of both lines of interpretation of the morphological record seems to be most likely.

During the shorter first phase of the climatic fluctuation, climate was glacier-friendly, whereas during the longer later phase, it favoured glacier recession and rock glacier development. This may have been caused by increasing dryness and a decrease in mean annual air temperature. Quantitative estimates of annual precipitation sums, based on ELA lowering and various glacier–climate models, suggest indirectly that summer temperatures were slightly lower (*c.* -1K to -0.5K) than during the 1931–60 period. Under such a scenario, precipitation was about 10–25% higher along the northern fringe of the Austrian Alps and about similar or slightly higher than today in the central areas. Even if we assume that summer temperatures remained unchanged (ΔT_s 0K), the necessary increase in precipitation of 25–30% along the northern–western fringe and of 20% in the central areas remains within realistic values. The lowering of annual temperature in the order of -1.5 to -2K agrees well with the value of -1.7K as given by von Grafenstein *et al.* (1998). Hence, a reasonable palaeoclimatic scenario points towards generally more humid conditions than today during the first part of the climatic fluctuation. During the second part, climate may have become drier with lower mean annual temperatures and increased seasonal temperature contrasts.

Further work will need to focus on the acquisition of more data from similar deposits in the vicinity of the Kromer valley itself and from other sites in the Alps, which had been correlated to the 'Kromer Stadial' in the past, to get a clearer picture of the timing of those glacier advances, which were clearly larger than 1850 (LIA), but smaller than the innermost 'Egesen Stadial' advances. This should also provide further insights into the early-Holocene glacier and climate history of the Alps.

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