Palaeoclimate records 60–8 ka in the Austrian and Swiss Alps and their forelands

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Abstract
The European Alps and their forelands provide a range of different archives and climate proxies for developing climate records in the time interval 60–8 thousand years (ka) ago. We review quantitative and semi-quantitative approaches for reconstructing climatic variables in the Austrian and Swiss sector of the Alpine region within this time interval. Available quantitative to semi-quantitative climate records in this region are mainly based on fossil assemblages of biota such as chironomids, cladocerans, coleopterans, diatoms and pollen preserved in lake sediments and peat, the analysis of oxygen isotopes in speleothems and lake sediment records, the reconstruction of past variations in treeline altitude, the reconstruction of past equilibrium line altitude and extent of glaciers based on geomorphological evidence, and the interpretation of past soil formation processes, dust deposition and permafrost as apparent in loess-palaeosol sequences. Palaeoclimate reconstructions in the Alpine region are affected by dating uncertainties increasing with age, the fragmentary nature of most of the available records, which typically only incorporate a fraction of the time interval of interest, and the limited replication of records within and between regions. Furthermore, there have been few attempts to cross-validate different approaches across this time interval to confirm reconstructed patterns of climatic change by several independent lines of evidence. Based on our review we identify a number of developments that would provide major advances for palaeoclimate reconstruction for the period 60–8 ka in the Alps and their forelands. These include (1) the compilation of individual, fragmentary records to longer and continuous reconstructions, (2) replication of climate records and the development of regional reconstructions for different parts of the Alps, (3) the cross-validation of different proxy-types and approaches, and (4) the reconstruction of past variations in climate gradients across the Alps and their forelands. Furthermore,
1. Introduction

Mountain regions provide excellent opportunities for developing palaeoclimatological records and approaches. The pronounced altitudinal and often also spatial climatic gradients allow the development and rigorous calibration of palaeotemperature proxy-indicators in archives as different as lake sediments and peat, speleothems, or tree-ring sequences (e.g. Lotter et al., 1997; Mangini et al., 2005). The landscapes in mountain regions are often strongly influenced by the steep climatic gradients and by past variations in climatic conditions. Therefore, the study of geological landscape features such as moraines, landslides and rock glaciers with appropriate geochronological approaches allows insights into past variations in climate (e.g. Ivy-Ochs et al., 2006; Kerschner and Ivy-Ochs, 2007). Other landscape features, such as the position of the Alpine treeline, are also strongly influenced by climate. Reconstructions of the past altitudinal limits of vegetation zones can, therefore, provide a further approach for constraining past variations in temperature (e.g. Tinner et al., 1996; Nicolussi et al., 2005).

In the European Alps and their forelands the earliest attempts to reconstruct past climate variations include interpretations of glacial landforms such as moraine ridges and erratic boulders as evidence of past glaciations (e.g. Venetz, 1833; Penck and Brückner, 1901/1909). Palynological (pollen-based) records and, more recently, macroscopic plant fossils preserved in lake sediment and peat have been used to make inferences about past vegetation and, indirectly, climatic change (e.g. Tinner et al., 1996; Lotter et al., 2000). In recent decades, stable isotope analyses and radiometric dating methods have broadened the range of archives that can be exploited for developing climate proxy records. Widely-used approaches in the European Alps and their forelands now include stable isotopic analyses in speleothems (e.g. Spötl et al., 2006; Boch et al., 2011) and lake sediment records (e.g. von Grafenstein et al., 1999; Lauterbach et al., 2011), dating of landforms such as terminal and lateral moraines as well as high-precision dating of organic matter deposited in glacier forefields (e.g. Joerin et al., 2006), analysis of fossil assemblages in lake sediment and peat to reconstruct past variations in biota and, indirectly, temperature (e.g. Heiri et al., 2004; Jost-Stauffer et al., 2005; Ilyashuk et al., 2011), and the study of past soil development to reconstruct phases of soil formation, aeolian sediment transport and permafrost (e.g. Terhorst et al., 2013).

Although palaeoclimate records from the Alpine region have a long tradition, they are often based on rigorous calibration studies, and have in many cases the potential for providing quantitative estimates of past climatic changes, they are presently underused in studies that examine climatic variations during the Last Glacial-to-interglacial cycle. A number of reasons may be responsible for this discrepancy. Many climate archives in the Alpine region are discontinuous over longer time intervals, since conditions ideal for accumulation of archives and suitable for recording climatic variations change both spatially and altitudinally during the examined time interval. Furthermore, dating accuracy and reliability varies dramatically in most archive types across the glacial–interglacial transition. As a consequence, many Late Pleistocene palaeoclimate records from the Alpine region do not have the resolution necessary for resolving leads or lags with other components of the climate systems, or with potential climate forcing factors. However, palaeotemperature or palaeoprecipitation signals may actually be recorded more reliably than in many other archives in adjacent lowland regions. In contrast to records produced in large, multinational projects, most palaeoclimatic reconstructions from the Alpine region are the products of collaborations between a small number of investigators or even single-author datasets. Although reconstructions are usually available and provided to interested colleagues upon request, many of the climate records from the Alps cannot be found in online climate data repositories. This is another reason why palaeoclimate records from the Alpine region may have been overlooked in synoptic studies of past climatic change.

The INTIMATE project aims at compiling proxy-based palaeoclimate records from Europe for the period 60–8 ka and facilitating the use of these data by the user community. This community includes climate modellers, who depend on independent climate records for evaluating model simulations, palaeoclimatologists developing and interpreting climate proxy records across Europe, and researchers studying the effects of past climatic changes on ecosystems and landscapes. Here we report the results of a workshop supported by INTIMATE which aimed at compiling and reviewing available palaeoclimate records from the Austrian and Swiss sectors of the Alpine region. The workshop, and therefore also the scope of this review, was intentionally restricted to Austria and Switzerland to reduce the complexity of the task and the number of records to be examined (Fig. 1). However, we also discuss a few selected records from eastern France and southernmost Germany since they help to constrain the past climate development in the northern Alpine forelands. Furthermore, it was decided to mainly focus on established proxies which allow quantitative, or at least semi-quantitative estimates of past variations in climate variables, since these reconstructions are directly comparable with other quantitative records of past climate change, climate model output data, and can be used as input variables to derive model based assessments of past ecosystem change (e.g. Heiri et al., 2006; Henne et al., 2011). However, an effort was made to cover a wide range of archives and palaeoclimatological methods, including palaeoecological, geochemical and geomorphological approaches.

We first provide an overview of the examined archives and proxy types (Section 2) and the available palaeoclimate records published for the study region within the time window 60–8 ka (Section 3). In a second step we discuss the spatial and temporal coverage of the presently available records and challenges faced by the community of researchers working with them. Finally, we list future steps necessary to increase the quality, accessibility and usefulness of palaeoclimatic records from the Alpine region in general, using the situation in Austria and Switzerland as an example. This review is complemented by parallel efforts by members of the INTIMATE group focusing on palaeoclimate data 60–8 ka available for Western Europe (Moreno et al., 2014), and Eastern Europe (Feurdean et al., 2014). As in all other attempts of inter-comparing independently dated late Quaternary records, chronological issues play an important role when compiling climate reconstructions in the Alpine region (Brauer et al., 2014). This is especially true if attempts are made to compare records from different archives, which are dated by different direct dating methods (e.g. radiocarbon, U/Th, annual
layer counting) or by indirect chrono-stratigraphical approaches. Age scales and reference ages which are presently used for late Quaternary records include the calibrated and uncalibrated radiocarbon age scale (0 = 1950 AD), the NGRIP Greenland ice core chronology ("B2k": 0 = 2000 AD), the BC/AD age scale, and different chronological techniques which consider the time of age determination or field sampling as a reference point. Here we refer to age estimates in thousands of years (ka) before 1950 AD, the zero point for the radiocarbon age scale. In addition, figures include the accepted time-scale for the presented records wherever this is practicable. Calibrated radiocarbon ages presented in the main manuscript text were calibrated using the IntCal13 calibration curve and CALIB 7.0.0 (Reimer et al., 2013) unless indicated otherwise. References to individual calibrated ages are reported as thousands of calibrated radiocarbon years BP (ka cal. BP), whereas ka is used for general age estimates which may be based on radiocarbon or other dating methods, or on a combination of approaches. To refer to climatic events in the Alpine region we use the regionally defined and accepted terms, which differ to some extent between archives (e.g. Table 1). The term “Lateglacial” refers to the common use in the Alpine realm. As used here it includes the Oldest Dryas Stadial (ending ~14.6 ka) and local glacier readvances, in contrast to the Pleniglacial, which refers to the period with extension of piedmont lobes of glaciers into the Alpine foreland.

Table 1
INTIMATE Greenland event stratigraphy (Blockley et al., 2012) and age of isotope events in the NGRIP ice core record (provided both as b2k and ka) compared with the established terminology for biozones, climate events, and glacial stages in the northern forelands of the Alps. Zones and events older than the Pleniglacial are not included (see Preusser, 2004 for a detailed compilation of older events and chronozones). Correlation of the Greenland event stratigraphy with biozones and glacial stages Preusser (2004). The correlation between biozones and glacial stages Preusser (2004). The correlation between isotope events in Greenland and biozones in the Alpine region is generally well established for the period <14.6 ka (e.g. van Radel et al., 2013), although there may be minor offsets in the timing and duration of biozones and climate events in Europe relative the Greenland isotope events they are correlated with (e.g. Rach et al., 2014). The correlation of Aegelsee Oscillation with the Older Dryas stadial follows Lotter et al. (1992).

<table>
<thead>
<tr>
<th>Greenland event stratigraphy</th>
<th>Age in NGRIP ice core (b2k)</th>
<th>Age (ka)</th>
<th>Biozones/climate events</th>
<th>Glacial stage</th>
</tr>
</thead>
<tbody>
<tr>
<td>Holocene</td>
<td>11,703-present</td>
<td>11.7-present</td>
<td>Holocene</td>
<td></td>
</tr>
<tr>
<td>GS-1</td>
<td>12,896–11,703</td>
<td>12.8–11.7</td>
<td>Younger Dryas stadial</td>
<td>Alpine Lateglacial</td>
</tr>
<tr>
<td>GI-1a, GI-1b, GI-1c</td>
<td>13,954–12,896</td>
<td>13.9–12.8</td>
<td>Allerød Interstadial (incl. Gerzensee Oscillation)</td>
<td>Egesen (glaciers shorter than Egesen)</td>
</tr>
<tr>
<td>GI-1d</td>
<td>14,075–13,954</td>
<td>14.0–13.9</td>
<td>Older Dryas stadial/Aegelsee Oscillation</td>
<td></td>
</tr>
<tr>
<td>GI-1e</td>
<td>14,692–14,075</td>
<td>14.6–14.0</td>
<td>Belling Interstadial</td>
<td></td>
</tr>
<tr>
<td>GS-2 and older</td>
<td>&gt;14,692</td>
<td>&gt;14.6</td>
<td>Oldest Dryas Stadial</td>
<td></td>
</tr>
</tbody>
</table>

Fig. 1. Location of records discussed in this article within the Alpine region. The symbols indicate different archive types, including lakes (blue triangles), palaeolakes and peat (black triangles), tree megafossils dated by dendrochronology (green circles), speleothems (red circles), moraines and glaciers (blue X symbols), and loess (green squares). A full list of localities is provided in the Supplementary data. The map was produced using stepmap (www.stepmap.de).
2. Examined archives and proxies

2.1. Lake and peat records

Lake sediments and peat deposits are common in the Alpine area, and can be analysed for geochemical and palaeoecological climate proxies. For the period 50–0 ka, these deposits can be reliably dated by radiocarbon dating of terrestrial plant remains. The age of older sediments has to be determined by, e.g., luminescence dating. In varved lake sediments, age estimates can be further constrained by annual layer counting (e.g. Brauer et al., 1999). Depending on the time interval and archive examined, estimated dating uncertainties can range from several years (some varved sequences in the Holocene) to a few centuries or even millennia (e.g. for some sequences older than 15 ka).

A wide range of palaeoecological, sedimentological and geochemical methods has been applied to lake and peat deposits to reconstruct past climatic change. Approaches that have hitherto been most successful in producing quantitative or semi-quantitative estimates of climatic changes 18–8 ka, and which therefore are reviewed here, include the analysis of the oxygen isotopic composition ($\delta^{18}O$) of lake sediment components (e.g. von Grafenstein et al., 1999; Lauterbach et al., 2011), the analysis of plant remains to reconstruct past altitudinal variations in the Alpine tree-line, providing evidence for variations in summer temperature (e.g. Tinner et al., 1996), and the analysis of fossil assemblages of pollen, or aquatic organisms such as chironomids, cladocerans, and diatoms (e.g. Lotter et al., 2000; Ilyashuk et al., 2011; Schmidt et al., 2012).

In central Europe, $\delta^{18}O$ of precipitation, and therefore also of lakewater, is strongly related to ambient air temperature (Rozanski et al., 1992; von Grafenstein et al., 1996). Since lakewater $\delta^{18}O$ determines $\delta^{18}O$ of sediment components produced within lakes, $\delta^{18}O$ records, e.g. from lacustrine carbonate, can be used to provide information about past temperature changes. However, water temperature directly affects and modifies the $\delta^{18}O$ signal recorded by carbonates produced in lakes, and this complicates the interpretation of carbonate $\delta^{18}O$ records (Von Grafenstein et al., 2000, 2013). Analyzing $\delta^{18}O$ of carbonate remains originating from the deep-water fauna of stratified lakes (e.g. ostracods), which remains unaffected by seasonal temperature changes, provides an avenue for alleviating this problem (e.g. von Grafenstein et al., 1999).

In the Alpine region, transfer functions have been widely used to develop palaeoclimate reconstructions from fossil pollen assemblages (e.g. Lotter et al., 2000), or fossil assemblages of aquatic biota such as chironomids, cladocerans, chrysophytes and diatoms (e.g. Schmidt et al., 2004, 2012; Ilyashuk et al., 2011). In Austria and Switzerland these inference models are typically calibrated based on the present-day relationship between these indicators and water temperature (e.g. Schmidt et al., 2012), air temperature (e.g. Lotter et al., 1997), or other seasonal aspects of the temperature regime (e.g. the date of autumn mixing; e.g. Schmidt et al., 2006).

The presence of forests in the northern Alpine forelands and the altitudinal position of the Alpine treeline provide important constraints for past summer temperature, since treeline elevation in the Alps is mainly controlled by temperature during the summer months. The interpretation of palaeobotanical records from many sites within a restricted geographical region allows the reconstruction of past treeline elevations (e.g. Nicolussi et al., 2005), the identification of short-term variations in treeline altitude (e.g. Lang, 1994; Tinner and Theurillat, 2003), and their interpretation as evidence for regional climatic oscillations (e.g. Haas et al., 1998). Reconstruction of past treeline altitude relies to a large extent on the identification, dating and analysis of tree remains in lake sediments and peat deposits, but also on fossil wood deposited in other environments (e.g. glacier forefronts, e.g. Nicolussi et al., 2005). At present, the minimum mean July air temperature for the growth of tree species in the Alps is ~7.5–9.5 °C (Landolt, 1992), and this threshold can be used to constrain past summer temperature based on variations in treeline altitude.

For the time interval 60–18 ka few quantitative estimates of past temperature based on lake and peat records are available. However, the presence of lakes during the last glaciation in itself provides valuable palaeoclimatological information, since lacustrine sedimentation requires ice-free conditions. Furthermore, the type and origin of lacustrine deposits (glaciolacustrine, organic lacustrine, peat), as well as plant remains such as pollen and plant macrofossils can provide additional information on the prevailing climate. For this interval we therefore also provide an overview and compilation of dated lake and peat deposits reported from Austria and Switzerland.

2.2. Speleothems

Cave carbonates are among the latest additions to the list of archives utilized to study palaeoclimate in the Austrian and Swiss Alps. First attempts to date Alpine speleothems occurred half a century ago (e.g. Franke, 1966), but modern palaeoclimatology research started only at the very end of the 20th century. A small number of data sets are currently available for the time interval of interest. These studies used stalagmites and flowstones from caves in Austria and southernmost Germany, and, more recently also from Switzerland. The chronology is exclusively based on the U–Th disequilibrium technique, which allows dating with an accuracy of typically a few hundred years for MIS 3 and better than a few decades for most of the Holocene (Brauer et al., 2014). This is of particular importance for the older part of the INTIMATE time interval where radiocarbon reaches its limit.

The main climate proxy used in speleothem studies is $\delta^{18}O$ of calcite, which reflects $\delta^{18}O$ of precipitation and, in the Alps, is positively correlated with mean annual air temperature (e.g., Darling, 2004). This allows the correlation of speleothem records to e.g. isotope records obtained from regional lake sediments and even to the ice cores from Greenland. Since $\delta^{18}O$ of water feeding speleothems is not only influenced by air temperature but also by changes in moisture sources, seasonality of precipitation, and locally important in-cave processes, translating speleothem $\delta^{18}O$ variations into quantitative estimates of past temperature change is challenging. Mangini et al. (2005) developed a quantitative reconstruction by calibrating $\delta^{18}O$ of a stalagmite from Spannagel Cave, Austria, against empirical constraints and estimates of past temperature for the calibration period 1688–1950 AD. This is currently the only speleothem record from the Alps (and for most of Central Europe) with a robust transfer function.

2.3. Glacier records

Glacier fluctuations are among the best-known and visible effects of climate variations. They represent an integrated signal of changes in accumulation and ablation, which can be parameterized as changes in summer temperature and precipitation. Due to the reaction time of glaciers, they tend to smooth out short-term
variations and respond to climatic change with decadal to centennial sensitivity depending on their size and topographic setting. Their use for continuous climate reconstructions is limited due to a number of reasons: First, moraines record only maximal extensions (usually cold phases), while remnants of short extensions overridden during subsequent readvances are destroyed and thus not preserved. Hence, the amplitude of glacier recession during warm phases can usually not be reconstructed. Earlier phases of glacial advances can be recorded by sediments, but the glacial record is almost always discontinuous, with gaps of unknown duration. Second, due to the paucity of organic material for radiocarbon dating in moraines or related sediments, glacial features could until recently almost only be dated indirectly by organic materials found in under- and/or overlying sediments and in some cases by radiocarbon ages from bones found in proglacial deposits. Cosmogenic Nuclides (e.g. $^{10}$Be) as well as Optically Stimulated Luminescence (OSL) and Infrared-stimulated Luminescence (IRSL) methods now allow the direct dating of some features, for example erratic boulders and glaciofluvial sediments, respectively. Thus, it has been possible to establish a framework of well-dated critical time points, but chronologies remain poorly constrained. Third, in most cases it is impossible to separate temperature and precipitation signals from the glacier record alone, and an independent proxy for at least one of the two variables is necessary. It is generally assumed that major glacier fluctuations in the Alps were mainly driven by variations in summer temperature, a parameter determining the energy fluxes towards the glacier surface, while regional amplitude differences are best explained by precipitation gradients (e.g. Penck and Brückner, 1901/09; Kerschner and Ivy-Ochs, 2007).

2.4. Loess-palaeosol sequences

Although largely absent within the Alps, loess-palaeosol sequences (LPS) in the Alpine forelands record variations in the broader region's palaeoclimate, especially phases of pedogenesis, permafrost and increased dust dynamics. Well-studied, high-resolution LPS are described from Austria (e.g. Haesaerts et al., 1996; Terhorst et al., 2013). No comparable records have been reported from Switzerland yet.

LPS are complex palsaeoenvironmental archives that form under variable climatic conditions by dust sedimentation in a dry tundra or (cold-)steppe environment (sedimentary aspect) and syn- to postsedimentary alteration from the surface of the deposits during moister and/or milder conditions and absence of significant dust input (pedogenic aspect). The degree of alteration in comparable parent material (loess) and relief position is a function of time and input (pedogenic aspect). The degree of alteration in comparable moister and/or milder conditions and absence of significant dust input is mainly controlled by dust sedimentation in a dry tundra and increased dust dynamics. Well-studied, high-resolution LPS in the Alpine forelands record variations in the climate of the heath. Their use for continuous climate reconstructions is limited

3. Overview of records

3.1. Lake sediments and peat

3.1.1. Lake and peat records 60–18 ka

There are a number of lacustrine and peat archives in the northern Alpine Foreland covering the early part of the Last Glacial cycle (Early Würmian, 115–74 ka; Preusser, 2004), such as Gondiswil, Füramoos, Samerberg, and Mondsee. Deposition of most of these sequences ended in the early Middle Würmian and limited information is available about environmental conditions during the Middle Würmian (74–30 ka) and the time when glaciers were present in the Alpine foreland (30–18 ka; Preusser, 2004). This lack of records can be explained by the filling of basins formed during the Rissian Glaciation, as well as by erosional processes during the Last Glaciation.

Key records, for which the age is constrained by geochronological data, originate from Niederweningen, Gossau and Huttwil-Galgenmoos (all northern Switzerland), and Unterangerberg and Baumkirchen (both western Austria) (Fig. 2). All of these records are floating, lacking continuous deposition until present. The deposits from Niederweningen (northern Switzerland) are known for their vertebrate remains, in particular of mammoth. The pedosol sequence is characterised by repeated changes of silty pond sediments and peat layers. Radiocarbon and luminescence dating indicate deposition of the peat around 45 ka (Hajdas et al., 2007; Preusser and Degering, 2007). Pollen, plant macrofossils, and beetle remains reflect environmental conditions dominated by open spruce–pine forest with larch and Swiss stone pine, mean summer temperatures between 8 and 13 °C, and mean winter temperatures between –20 and –5 °C (Coope, 2007; Drescher-Schneider et al., 2007). Sediments exposed in the gravel pit of Gossau show several alternating layers of sandy-siltly gravel, silt and peat. Radiocarbon, U-series and luminescence dating constrain the sediment age to 60–30 ka (Preusser et al., 2003). The pollen record reflects open pine–spruce forest with Swiss stone pine during the older peat layer, and open pine forest during the younger peat. The beetle fauna in the younger layer indicates mean summer temperatures between 8 and 12 °C and mean winter temperatures between –21 and –14 °C (Jost-Stauffer et al., 2001). For the lower peat, a mean summer temperature of 12–13 °C, and a mean winter temperature between –15 °C and –7 °C have been reconstructed (Jost-Stauffer et al., 2005).

Sediments from a palaeolake in Unterangerberg revealed a complex lithology comprising diamic, gravel, sand, lignite and thick intervals of fine-grained sediments. Beside sediments older than ~100 ka, the sequence recorded the basin development from the end of the Early Würmian (MIS 5a) up to ~40 ka. Pollen assemblages indicate grassland vegetation with scattered stands of pine during MIS 4 and parts of MIS 3. An interstatal dominated by birch is dated to ~50 ka, a second one with open pine–spruce forest to ~45 ka (Starnberger et al., 2013).

The record from Baumkirchen comprises a >200 m sequence of laminated lake sediments. Uncalibrated radiocarbon ages from 25.5 to 32.4 ka (e.g. Pli et al., 1971, 1972) originally suggested a sedimentation age for a 26 m-thick section between ~29 and 36 ka. Re-dating of several samples using AMS resulted in a revised age of ~35–36 ka cal. BP (Spötl et al., 2013). Luminescence dating yielded mean ages of 37.3 ± 3.5 ka (IRSL UV emissions) and 37.8 ± 3.6 ka (post-IR OSL UV emissions), whereas IRSL (blue emissions) gave a higher mean age of 45.0 ± 4.3 ka (Klassen et al., 2007). A palynological study of an 86 cm section revealed 17 vegetation periods representing grassland tundra with few birch and pine trees (Bortenschlager and Bortenschlager, 1978).
indicates deposition of the sandy-silty slope deposits ~50 ka, chronological position controversial, cf.; Preusser, 2004). IRSL dating based on ostracods, molluscs, and charophyte remains (Von Grafenstein et al., 1999, 2013; Lauterbach et al., 2011). In the Alpine region, these short-term oscillations are not consistently apparent in summer temperature reconstructions based on biotic proxy records, such as Jeserzersee and Lauterbach et al., 2011). Lotter et al. (2012) discuss possible reasons for the discrepancy in temperature trends reconstructed by pollen and chironomid data from the Swiss Plateau (Von Grafenstein et al., 1999, 2013). Diverging centennial-to-millennial-scale temperature trends are reconstructed during the Bolling-Allerød period (~14.6–12.8 ka) by different proxy types. Chironomid-based reconstructions tend to infer no trend or a gradual warming (e.g. Heiri and Millet, 2005; Lotter et al., 2012; Samartin et al., 2012a, b). A pollen-based reconstruction from Gerzensee indicates decreasing summer temperatures (Lotter et al., 2012). Reconstructions of δ18O of precipitation, finally, indicate no distinct centennial- to millennial-scale trend in mean annual temperature (Von Grafenstein et al., 1999, 2013; Lauterbach et al., 2011). Lotter et al. (2012) discuss possible reasons for the discrepancy in temperature trends reconstructed by pollen and chironomids at Gerzensee and suggest that it may be explained by the influence of winter temperatures or precipitation on vegetation on the Swiss Plateau. Tree macrofossil and stomata data from the Bernese Alps and their foreland (Fig. 4) suggest that treeline elevation in the northern Alps may have reached 700–1300 m a.s.l. during the Bolling-Allerød period. Paleobotanical records from the Swiss Central Alps and the southern Alps document that treeline during the Bolling-Allerød period had probably already reached elevations around 1600–2000 m a.s.l. (Welten, 1982; Lang and Tabor, 1989; Tinner and Vescovi, 2007).

δ18O records provide evidence for decadal to centennial-scale cooling episodes at ~14.3 ka (Lotter et al., 1992; von Grafenstein et al., 1999, 2013; Lauterbach et al., 2011). In the Alpine region, these short-term oscillations are not consistently apparent in summer temperature reconstructions based on biotic proxy records, such as Jeserzersee and Lauterbach et al., 2011). Lotter et al. (2012) discuss possible reasons for the discrepancy in temperature trends reconstructed by pollen and chironomids at Gerzensee and suggest that it may be explained by the influence of winter temperatures or precipitation on vegetation on the Swiss Plateau. Tree macrofossil and stomata data from the Bernese Alps and their foreland (Fig. 4) suggest that treeline elevation in the northern Alps may have reached 700–1300 m a.s.l. during the Bolling-Allerød period. Paleobotanical records from the Swiss Central Alps and the southern Alps document that treeline during the Bolling-Allerød period had probably already reached elevations around 1600–2000 m a.s.l. (Welten, 1982; Lang and Tabor, 1989; Tinner and Vescovi, 2007).

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remains, perhaps due to their relatively small amplitude, the discontinuous sampling in many records, and the lag which may be expected before biological indicators respond to climatic change. Alternatively, the temperature decrease during these events may have been less prominent during summer than during the rest of the year. In records where they are apparent, the summer temperature decrease during these events has been estimated to ~0.5°C to ~3°C (e.g. Lotter et al., 2000; Ilyashuk et al., 2009; Lotter et al., 2012).

A distinct and rapid cooling in summer temperature of ~1.5–3°C is reconstructed based on pollen, chironomid, and cladoceran records at the beginning of the Younger Dryas cold episode (~12.7 ka), followed by a phase of rapid warming by ~1.5–4°C at the beginning of the Holocene at ~11.7 ka. These two temperature transitions were accompanied by rapid shifts in δ18O of precipitation of ~2–2.5‰ in the northern forelands of the Alps (von Grafenstein et al., 1999, 2000; Lauterbach et al., 2011). Treeline declined during the Younger Dryas (Tobolski and Ammann, 2000;
however the altitudinal extent of this decline is not clear. Decreases in treeline elevations were apparently relatively minor in the northern Alps (e.g. Fig. 4A) but may have reached 300–400 m altitude in the southern part of the Swiss Alps (Tinner and Vescovi, 2007). Temperature reconstructions based on lake sediments covering large intervals of the Holocene include the Ammersee 18O record, and several transfer function-based reconstructions from fossil assemblage data (e.g. Egelsee: Larocque-Tobler et al., 2009; Hinterburgsee: Heiri et al., 2003a, 2004; Schwarzsee ob Sölden: Wick, 2000), however the altitudinal extent of this decline is not clear. Decreases in treeline elevations were apparently relatively minor in the northern Alps (e.g. Fig. 4A) but may have reached 300–400 m altitude in the southern part of the Swiss Alps (Tinner and Vescovi, 2007).

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Ilyashuk et al., 2011). Furthermore, a number of plant macro- and megafossil records provide constraints for the maximum altitude of past treeline elevations in the Swiss and Austrian Alps (e.g. Tinner and Theurillat, 2003; Nicolussi et al., 2005). Temperature variations of the Holocene are considerably more difficult to track based on lake sediment records than the higher-amplitude variations of the Lateglacial period. Reconstructions of δ¹⁸O of precipitation suggest a number of centennial-scale oscillations (e.g. the Preboreal Oscillation 11.3–11.4 ka; von Grafenstein et al., 1999; Lauterbach et al., 2011), although with the exception of the cooling at 8.2 ka, their amplitude is lower than even minor oscillations recorded during the Lateglacial. Palaeobotanical records from sites near the Alpine treeline similarly provide evidence for centennial-scale temperature decreases (e.g. Haas et al., 1998; Tinner and Theurillat, 2003). Reconstructions from Austria and Switzerland based on fossil assemblages of aquatic biota also document centennial-scale cooling episodes during the Holocene (e.g. Heiri et al., 2003a, 2004; Larocque-Tobler et al., 2009; Ilyashuk et al., 2011). However, not all the oscillations which apparently took place are visible in each of the available records, which may be related to their small amplitude in comparison with the errors of the applied transfer functions (typically 1–1.5 °C; e.g. Lotter et al., 1997; Heiri et al., 2011), but also to discontinuous sampling, or regional differences in the amplitude of these events across the region. Furthermore temperature variations may have been less extreme in summer than during other seasons. Centennial-scale temperature changes recorded during the recent past in Switzerland (i.e. the temperature rise during the 19th—20th century) were less pronounced during the summer months than during winter and autumn (Begert et al., 2005; Rebetez, Reinhard, 2008).

After the Younger Dryas—Holocene transition, treeline altitude increased and reached maximum elevations in the Austrian and Swiss Alps within the period ~10–5 ka (e.g. Lang, 1994; Tinner and Theurillat, 2003; Nicolussi et al., 2005; Lotter et al., 2006, Fig. 3). The treeline rise in the Bernese Alps was at least 700 m between the Younger Dryas and Holocene optimum, from below 1400 m to over 2100 m a.s.l (Fig. 4A). Similar estimates of treeline rise by about 700–800 m at the onset of the Holocene are available from the Central Swiss and Southern Alps (Tinner and Kaledriened, 2005; Tinner and Vescovi, 2007). Transfer function-based reconstructions from fossil assemblage data also indicate a relatively warm early and mid-Holocene in Switzerland, with highest mean July air temperatures inferred within the period ~10–6 ka (e.g. Heiri et al., 2003a, 2004; Larocque-Tobler et al., 2009), although the exact timing of the warmest phase of the Holocene differs between records. The only available transfer function-based July air temperature record from the Austrian Alps is from the Kleegruben Cave (~2165 m a.s.l) whose top part grew from nearby Spannagel Cave (~2500 m a.s.l.) whose top part grew between 60 and 53 ka. However, the very slow growth rate (~6 mm/ka) and the possible presence of hiatuses render a detailed correlation with the Kleegruben Cave record difficult (Holzkämper et al., 2005). The magnitude of the isotopic stadial-interstadial changes, however, is comparable to the Kleegruben stalagmites (~2‰).

New records will soon become available from Hölloch (Allgäu Mountains), Germany’s currently second largest cave straddling the border to Austria. Preliminary data from one stalagmite show δ¹⁸O jumps of 3‰ at the onset of interstadials between GIS 17 and 11 (Spötl et al., 2011).

Speleothem growth during glacial maxima is unlikely given the lack of soil and vegetation in the catchment of the cave’s drip waters. Studies in Spannagel Cave demonstrated, however, that calcite deposition is locally possible when the cave is overlain by ice (Spötl and Mangini, 2007). Three conditions have to be met to allow this rare mode of speleothem formation: first, the presence of sulphide minerals in the aquifer giving rise to corrosive waters upon weathering (driving karst dissolution in the absence of carbon dioxide-enriched waters from soils); second, the presence of a temperate glacier above the cave preventing the karst aquifer from freezing; and, third, galleries that are not flooded by glacier melt waters.

3.2.2. Speleothem records 18–8 ka

A stalagmite was retrieved from Hölloch (Allgäu Mountains), which started growing in the Allerød and includes the Holocene, probably compromised by short hiatuses (Wurth et al., 2004). The most prominent isotope excursion is the 2% drop in δ¹⁸O coincident with the Younger Dryas (Fig. 5), similar in amplitude as in ostracod δ¹³C records from Ammersee (von Grafenstein et al., 1999) and Mondsee (Lauterbach et al., 2011). The published chronology of the Hölloch stalagmite is based on eight U–Th dates and was later adjusted due to a spike correction (Niggemann, 2006).

A few other stalagmites from the Eastern Alps recorded the Allerød—Younger Dryas—Holocene transitions by δ¹³C drops consistent with the Hölloch stalagmite. These include, e.g., a stalagmite from Blasloch, Styria (Spötl et al., 2007).

Two sites have been studied for early Holocene speleothem growth in the Eastern Alps, Spannagel Cave and Katerloch (20 km NE of Graz, Austria), and one site in the Western Alps (Milchbach Cave, a site next to the retreating Upper Grindelwald Glacier, Switzerland). Three stalagmites from Spannagel Cave were combined to a stack that comprises the last 9 ka of the Holocene (Vollweiler et al., 2006), recently extended to 11 ka using a total of five stalagmites (Fohlmeister et al., 2013). The most recent part of this record (the last 2 ka) was calibrated as a temperature record using empirical constraints (Mangini et al., 2005).

Katerloch stalagmites yielded the first highly resolved (average 1.5–4 ka) and well-dated proxy record of the 10.0, 9.1 and 8.2 ka events in the Alps (Boch et al., 2009). These cold spells are expressed by negative δ¹⁸O anomalies of up to 1.1‰, similar in magnitude to the excursions recorded in the δ¹⁸O record from Ammersee (von Grafenstein et al., 1999).
A series of stalagmites from Milchbach Cave (Luetscher et al., 2011) document continuous calcite growth since 9.3 ka. Changes in the calcite fabric as well as in the stable isotopes reflect the waxing and waning of the nearby glacier, which affected soil development, erosion, and cave ventilation regime. Although detailed stable isotope profiles still await publication, the data document an extended period of glacier retreat between ~9.3 and 6.8 ka, interrupted only by short advances, the most significant of which occurred at 8.2 ka (Luetscher et al., 2011).

3.3. Glacier records

3.3.1. Glacier records 60–18 ka

The gravel and peat section of Gossau, Switzerland (Preusser et al., 2003; see Section 3.1.1), indicates that the Linth glacier did not reach the foreland of the Alps between ~60 and 30 ka. Similar evidence exists for the Inn valley at Unterangerberg, Austria (Starnberger et al., 2013), and the site of Baumkirchen indicates that the advance of ice into the lower Inn valley occurred subsequent to 32–33 ka (Spotl et al., 2013).

For the last glaciation of the foreland (the “Alpine LGM”), the best constrained data set is available for the Rhine-Linth glacier system, which formed two coalescent piedmont lobes over northern Switzerland and southernmost Germany (data compiled by Keller and Krayss, 2005a, b; re-assessed and calibrated by Schoeneich, 2011). The available data document the progression of the piedmont lobes after ice-free conditions, some oscillations during the phase of maximal extent, and minor readvances during the early recessional phase. Detailed mapping of moraines and melt-water channels allowed the reconstruction of four main stages of the piedmont lobes, known as the Schaffhausen/Killwangen, Feuerthalen/Schlieren, Stein am Rhein/Zurich and Konstanz/Hurden stadials (e.g. Keller and Krayss, 2005a) (Fig. 6). The first three stages were formed by glacier advances, followed by several recessional moraines, whereas the last stage represents a phase of ice stagnation.

Eight radiocarbon dates on peat layers covered by till from sites situated between the Alpine front and Lake Constance show that the progradation of the glacier lobes over the foreland started not earlier than 32 ka. The maximal extent at Schaffhausen was reached after 30 ka (Frank and Rey, 1996). At Knollengraben, between the positions of the Stein am Rhein and Konstanz stadials, a peat layer between two tills records an oscillation dated to 26.0–27.5 ka. It might represent a retreat preceding the Stein am Rhein readvance (Weinhold, 1973), but was later reinterpreted as an oscillation during a glacier advance before the maximum (Keller and Krayss, 2005b). OSL dating of melt-water sediments (Preusser et al., 2007) and a comparison with dates from the Linth lobe (Schlüchter and Röthlisberger, 1995) favour the first hypothesis, suggesting a maximal advance between 30 and 27 ka. The presence of glaciers in the foreland is also documented for the northern lobe of the Valais glacier by OSL and radiocarbon dating of the Finsterhennen site (Preusser et al., 2007). Here, the maximum extent is dated by 10Be to ~24 ka (Ivy-Ochs et al., 2008; recalculated after; Balco et al., 2009). Thus, the advance of the northern Alpine piedmont lobes to the furthest extent may have occurred already during GS-5 and/or GS-4, with a second advance when the glaciers were quasi-stationary for a longer period of time during GS-3. Minimum ages for glacier retreat are given by some organic layers overlying till and by palaeomagnetic secular variations in both basins of Lake Constance and in Lake Zurich (Niessen et al., 1992; Wessels, 1996). Measurable secular variations can only be recorded in very fine and calm sedimentation conditions, excluding turbidites, dropstones and any kind of high energy sediments, thus indicating totally glacier free lakes. Accordingly, Lake Constance and Lake Zurich were ice-free no later than ~17–18 ka, but possibly before 20 ka. This result is consistent with the deepest radiocarbon date from the Lake Zurich cores, which yielded an age of 14600 ± 250 14C yr.
(Lister, 1988) equivalent to –18.4–17.1 ka cal. BP. AMS radiocarbon dates on terrestrial plant macrofossils (Ericaceae and Juniperus wood) show that smaller lakes in the piedmonts of the northern Swiss Alps (e.g. Wauwilermoos) became ice-free before 17.7–18.6 ka cal. BP, when several meters of fine-graded lacustrine sediments had already accumulated on the top of the LGM moraine material (Beckmann, 2004).

The successive piedmont stages of the Rhine-Linth glacier were reconstructed cartographically (Keller and Krayss, 1994, 2005a). Volume calculations based on these maps (Benz-Meier, 2003; Keller and Krayss, 2005b) show that 70% of the ice volume had melted before 20–22 ka, and 95% of the volume was gone before the Gschnitz stadial (Fig. 6).

### 3.3.2. Glacier records 18–8 ka

The Lateglacial in the Eastern Alpine valleys starts with a poorly constrained period during which the main glaciers in the large longitudinal valleys experienced downwasting and stagnation (Reitner, 2007), whereas smaller, already disconnected glaciers may have experienced a few phases of glacier growth (van Husen, 1997). In the main valleys of the Eastern Alps large systems of ice marginal terraces document the downwasting of the dendritic glacier systems. This phase of the early Lateglacial ice decay is centred on ~19 ka (Klasen et al., 2007), the end is marked by the radiocarbon date of Rööschitz, Austria (18.0–19.1 ka cal. BP, van Husen, 1997).

The first well-defined re-advance is known as the Gschnitz stadial (Penck and Brückner, 1901/09). At the type locality south of Innsbruck (Tyrol, Austria), stabilization of the moraines of a local glacier is dated to 16–17 ka (10Be dates, Ivy-Ochs et al., 2006, recalculated after Balco et al., 2009). Moraines attributed to the Gschnitz stadial are present in other valleys as well. The glacier advance is interpreted as a reaction to the Heinrich 1 ice-rafting event in the North Atlantic. The Gschnitz stadial was followed by a series of glacier oscillations until the onset of the Bolling-Allerød Interstadial. In the Eastern Alps they are known as the Clavadel/Senders and Daun stadials (Maisch, 1981).
In the Swiss Prealps, an exceptionally well-preserved moraine sequence records high-frequency oscillations of small glaciers (0.5–4 km length) (Schoeneich, 1998a, b). The sequence contains up to 6 “stadials”, each containing 3 to 5 minor oscillations, with a total of over 30 oscillations bracketed between −19 and 15 ka. Three generations of rock glaciers are found between stadials. A comparison with historical fluctuations of glaciers of similar size shows that the main advances were of similar amplitude as the main LIA oscillations, and that minor oscillations were in the range of recent decadal oscillations (1890, 1920, 1970’s). This suggests that a centennial to decadal variability of similar amplitude as historical variations existed throughout the early Lateglacial.

During the Younger Dryas event, a multi-phase readvance, the so-called Egesen stadial, is well-documented in numerous valleys, and dated at several places in the Alps (Ivy-Ochs et al., 2009; Schindelwig et al., 2011). Field evidence and datings point to at least two separated advance phases with minor oscillations in between. Available dates (recalculated after Balco et al., 2009) suggest that the first and main advance coincides with the first few centuries of the cold event, and the second peak to the second half of the Younger Dryas. Rock-glacier activity at lower altitudes within the glacier-covered areas of the Egesen maximum advance ended at the beginning of the Holocene (Hofmann et al., 2006; Schoeneich, 2009).

For the palaeoclimatic interpretation of glacier records 18–8 ka, the equilibrium line altitude (ELA) is the most useful starting point. There, accumulation equals ablation, which can be parameterized by precipitation and summer temperature (Ohmura et al., 1992). For studies in the European Alps, the ELA is usually calculated from a map of the glacier surface using an accumulation area ratio (AAR) of 0.67. ELA fluctuations are reported as differences to the Little Ice Age reference level (ELA depressions) (Gross et al., 1977). For climatic inferences, they must be adjusted to a well-defined standard climate period, such as 1961–90 (Kerschner and Ivy-Ochs, 2007). ELA depressions for the Gschnitz stadial in the central Alps are in the order of 600–700 m. In the Northern Alps, scarce data point to values in the order of 800 m, while in the southern Eastern Alps they range around 900 values in the order of 800 m, while in the southern Eastern Alps/C14 of 10

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Loess research has a long tradition in Quaternary stratigraphy, but only recent advances in geochronology and the development of high-resolution datasets (e.g. based on magnetic susceptibility, granulometry, geochemistry) have allowed the reconstruction of palaeoclimatic records for different loess regions. As a consequence, first attempts are now made to correlate Late Pleistocene LPS with e.g. ice core records (e.g. Antoine et al., 2009; Haesaerts et al., 2010; Terhorst et al., 2013). Due to high sedimentation rates, the LPS Nussloch near Heidelberg, Germany, is regarded as Central European reference sequence. Based on grain size variations, a correlation with the GRIP dust record has been proposed (Antoine et al., 2009). Here we discuss two well-resolved LPS in eastern Austria (Wachtberg, Willendorf) as examples to demonstrate the type of information about past climatic change that may be developed from loess deposits in the northern Alpine forelands.

The age of the LPS of Willendorf (Fig. 7A) with numerous Upper Palaeolithic cultural layers is relatively well constrained by radiocarbon dating (Haesaerts et al., 1996, 2013; Nigst et al., 2008), thus enabling the compilation of a preliminary age model. Unit E (loess) and units below are not dated yet. Unit D consists mainly of cambic horizons that formed most likely during early to middle MIS 3 (Haesaerts et al., 1996). Mollusc assemblages of mixed forest ecosystems in unit D confirm a climate mild and moist enough for forest development, reflecting the Willendorf interstadial (Haesaerts et al., 1996). An intercalated loess unit (D3; Nigst et al., 2008) represents a cold–dry phase, possibly correlating to Heinrich Event 5. Unit C is a complex sequence of loess and soil horizons, representing on average cooler climate with significant oscillations, including phases of permafrost. Three humic palaeosol horizons are attributed to the Schwallenbach interstadials I to III. This period ended at 31.4–34.7 ka cal. BP (Haesaerts et al., 1996). With the exception of a basal bleached (B4) and one weak humic horizon (B2), Unit B consists of loess, which developed during cold–dry periglacial conditions (27.2–27.7 ka cal. BP; Haesaerts et al., 1996).

At Krems-Wachtberg (Fig. 7C), 24 km downstream of the Danube, an 8 m thick LPS formed between –40 and 22 ka. Redeposition and erosion was weak at this locality, as indicated by largely continuous luminescence datings (Lomax et al., 2012; Händel et al., 2013). Radiocarbon dates are only available from the archaeological horizons (AH) (full list of calibrated ages provided in the supplementary information). The main AH age dates to 31–33 ka cal. BP (Simon et al., 2013). The lower part of the sequence (GH 37–23) with AH 4 in its upper part formed before 30 ka and is characterised by bleached and incipient soil horizons. The absence of well-developed cambic horizons in the upper part of the LPS
indicates increasing sedimentation rates and overall colder climate. A chronological framework for the whole sequence, based on palaeoepedological, magnetic, and chemical analyses is suggested by Hambach et al. (2008) and further developed by Terhorst et al. (2013).

Palaeoclimatic indications and trends derived from these high-resolution records are supported by data derived from studies on less resolved LPS in the Austrian loess belt. LPS from Stratizing and Langenois, some km NE to Krems-Wachtberg, contain tundra gley soils that developed during MIS 3 (Thiel et al., 2011), but lack a detailed chronology. Stillfried B in East Austria is a former type section for the Middle- to Late Würmian transition, but this sequence is not consistently dated yet (Terhorst et al., 2011). Poly-pedogenesis and redeposition hamper unambiguous correlations to other records. In the northwest Austrian Alpine forelands, closer to the Alpine margin, LPS are characterized by well-developed pedocomplexes from the MIS 3, most likely due to higher average precipitation compared to the sites further east (Terhorst et al., 2002). First OSL datings indicate phases of advanced weathering, interrupted by permafrost during MIS 3, whereas MIS 2 is represented by a series of loess and tundra gleys (Terhorst et al., 2002).

4. Discussion

The archives and proxy types reviewed in Section 3 document the wide range of approaches used in the Alpine region to quantify past climatic variations 60–8 ka. We restricted this review both geographically (Switzerland, Austria, and some adjacent regions) and in respect to the examined proxies. The main focus was on established approaches providing quantitative palaeoclimatic estimates, although, due to the scarcity of records, a number of non-quantitative approaches were discussed for the time window 60–18 ka as well (e.g. Sections 3.1.1, 3.4). Without these constraints the list of available approaches could have been extended considerably. Furthermore, for several proxy types we present case studies and examples suitable for demonstrating the potential of a...
given proxy (e.g. in respect to treeline changes, glacier fluctuations, or LPS; Sections 3.1.3–3.4), rather than a full overview of all available records. Nevertheless, the discussed examples provide an overview of the state of the art of quantitative palaeoclimate reconstruction 60–8 ka in the Austrian and Swiss Alps and of the steps necessary for further advancements in the field.

4.1. Present situation and issues in respect to existing records

4.1.1. Dating and hiatuses

Even a coarse examination of the different records and archives presented here illustrates the large variation in dating accuracy and resolution between archives. For example, lake sediment records from the Alpine region, even non-varved sequences, can usually be dated with a relatively high accuracy for the early and mid Holocene (e.g. Ilyashuk et al., 2009, Larocque-Tobler et al., 2009, Lauterbach et al., 2011). Dating of sediments deposited 14–11 ka is more problematic, whereas beyond ~14.6 ka dating uncertainties typically increase dramatically (e.g. Huber et al., 2010; Schmidt et al., 2012). The situation is similar for records of past glacier extent. During the Holocene, glaciers expanded several times to altitudes below treeline. Plant macrofossils for radiocarbon dating, that can be used to support and cross-validate other applicable dating methods, are therefore relatively common in glacial deposits and glacier forefields (e.g. Joerin et al., 2006; Nicolussi and Schlüchter, 2012). In older deposits terrestrial plant material tends to be sparse, and dating typically relies on luminescence or on combined approaches, a similar situation as encountered when dating Late Pleistocene loess sequences. Errors associated with all of these approaches inevitably increase in older deposits. Several records are described for which different dating approaches converge, leading to robust and independently validated age estimates even for relatively old deposits (e.g. Starnberger et al., 2013, Fig. 7C). Older records within the time window 60–8 ka are more often affected by hiatuses, either because of discontinuous accumulation (e.g. lakes, speleothems, loess) or because erosive processes triggered by climate change destroyed previously deposited evidence (e.g. glacier advances).

The problems outlined above are to a large extent inherent to late Quaternary deposits, and not restricted to the Alpine region. Since landscapes and deposition processes change with changing climates, the availability of records and the suitability of deposits for different dating approaches change as well. However, several of the problems can be addressed or alleviated to some extent if research projects are designed accordingly. For example, lake sediment studies aiming to obtain larger amounts of sediment (e.g. using larger-diameter corers or replicate coring) will have a higher chance of isolating terrestrial plant remains suitable for radiocarbon dating. Alternatively, newly developed approaches for dating late Quaternary sequences, e.g. compound or fossil-specific radiocarbon dating (e.g. Tennant et al., 2013) or microfert analyses (e.g. Pollard et al., 2006; Huber et al., 2010), may significantly reduce dating uncertainties for the time interval 50–15 ka. For glacier studies extensive and structured searches for radiocarbon-datable material in glacial deposits or glacier forefields may yield additional age constraints. Further efforts to establish independently cross-validated chronologies based on several independent dating methods are needed to better constrain palaeoclimatic records >15 ka. This may allow the development of new proxy data even for time intervals presently without or with very few proxy-based constraints on palaeoclimatic conditions. Recent developments in speleothem research are promising in this respect. New records are now emerging that cover the Alpine LGM (Luetscher and Spötl, unpublished) even though this period has previously been considered to be poorly suited for speleothem growth.

4.1.2. Fragmentary records and replication (temporal coverage)

Proxy records in the Alpine region 60–8 ka are not only affected by hiatuses for some intervals, but also by a relatively short record length and a low degree of replication. For example, several chironomid-based temperature reconstructions are available for Switzerland, making this one of the most replicated approaches for quantitative climate reconstruction based on lake sediments in this region (see Section 3.1.2 and Fig. 3). A number of records are available that cover the transition from the earliest Lateglacial to the early Holocene (~15–11 ka), several that extend from the Younger Dryas to the mid Holocene (~12–8 ka), and a few which extend back to the earliest Lateglacial (~17 ka). However, no record is presently available which encompasses the entire time window 17–8 ka. Temperatures in some sections of this interval (e.g. 15–17 ka) are only constrained by a single dated chironomid record. For other quantitative temperature proxies from lake sediment records, temporal coverage and replication are even sparser than for chironomids. A similar situation is apparent for speleothem records: Variations in δ18O have only been replicated for a few time intervals and regions (e.g. Spotl et al., 2006; Vollweiler et al., 2006). Large parts of the time period 60–8 ka are, however, represented only by a single speleothem record in the Swiss and Austrian Alps. A better replication of such records within specific sectors of the Alps would allow a more rigorous evaluation of the reliability and regional relevance of individual records (Lotter et al., 1992). For example, a recent record of δ18O of ostracod valves from Mondsee, Austria (Lauterbach et al., 2011), shows an excellent agreement with an earlier record from Ammersee in the Alpine foreland of southern Germany (von Grafenstein et al., 1999), as well as with stalagmites from Höllloch Cave (Wurth et al., 2004), supporting the reliability of decadal- to millennial-scale variations recorded in these records. However, this comparison also reveals that a prominent excursion recorded in the Ammersee record within the Younger Dryas cold phase ~12 ka is not represented in the other records, and this event is clearly not representative for the regional climate development.

Some disciplines, such as glaciology and palaeobotany, have a longer tradition in the Alpine region than studies of aquatic proxies in lake sediments or speleothems, and a comparably high number of data and studied sites is available for inferring past glacier variability and vegetation change. However, to develop records representing long-term variations in climatic variables, glaciological and palaeobotanical datasets have to be compiled to regional reconstructions. Here, similar difficulties are apparent as with lake sediments and speleothem records. Few regional reconstructions are available that cover large sections of the interval 60–8 ka. For example, in the Bernese Alps and their foreland a considerable number of plant macrofossil records is now available, each providing information on the presence or absence of tree vegetation at a certain altitude during sections of the Lateglacial period or the Holocene (see Fig. 4A). To produce a continuous reconstruction of past variations in treeline these records have to be assembled and compared, allowing inferences to be made about the altitudinal limits of past tree growth. This information can, in turn, be used to produce estimates of past summer temperature change in the Alps (see, e.g. Fig 4B). Both the overall altitudinal development of treeline within the region and the gaps in the records are not apparent to non-specialists if the data are not compiled into overview diagrams and simplified datasets. A similar situation is apparent regarding loess records and reconstructions of past glacier extent. Without visualising and summarizing the complex stratigraphical and chronological data, often from different study sites (e.g. as done
in Figs. 6–7), the records are difficult to interpret for researchers unfamiliar with these approaches.

4.1.3. Regional and local effects (spatial coverage)

Regional differences in the efforts invested into developing and replicating proxy records also make it difficult to identify large-scale patterns in climate development across the Alpine region and separate them from local artefacts. As an example, a recently published chironomid record from Schwarzeise b Sölden in the Eastern Alps (Ilyashuk et al., 2011) indicates relatively high early Holocene temperatures at the study site compared to chironomid-based reconstructions and treeline records from the northern Swiss Alps (e.g. Heiri et al., 2004 and Fig. 4B). Confirmation of this difference by additional chironomid records is still lacking, and therefore it is unclear whether this finding reflects a true difference in regional temperature development, whether it reflects exceptional early Holocene temperatures at a particular site, or whether the biotic response to climate was amplified due to site-specific conditions. As discussed in Section 3.3.2., regional studies of ELA provide indications for changing precipitation gradients within the Alps during the Younger Dryas. More systematic replication and spatial coverage of proxy records across the Alps would allow similar comparisons to be made for other proxies to detect regional differences in proxy response, and reconstructed temperature development. For example, some of the available palaeobotanical records from around treeline elevation suggest that the Alpine treeline reached maximum elevations earlier in the Central Alpine region (10–5 ka; Tinner et al., 1996) and the Eastern Alps (>6.5 ka; Nicolussi et al., 2005) than in the northern Alpine region (6–4 ka; Lotter et al., 2006), suggesting regional differences in the timing of the Holocene maximum in summer temperature across the Alps.

4.1.4. Quantification and validation of proxies

For some proxy types, major progress has been made in recent years in quantifying the relationship between proxy indicators and the environmental variable(s) they represent. These advancements typically involve the development of empirical or mechanistic numerical models that allow quantitative inferences about past climatic conditions based on proxy data. An example of this approach is the development of transfer functions based on fossil assemblages from lakes covering a wide range of environmental conditions (calibration in space; e.g. Lotter et al., 1997; Heiri et al., 2011). Alternatively, well-dated proxy sequences can be related to instrumental or observational time series (calibration in time; e.g. Mangini et al., 2005). An example of mechanistic models applied for reconstructing past climatic changes in the Alpine region is provided by reconstructions based on ELA changes. If assumptions are made regarding the remaining variables influencing glacier mass balance, ELA variations can be used to infer past variations of summer temperature or precipitation (e.g. Ivy-Ochs et al., 2006 and Section 3.3.2.). Most of these approaches have in common that they have been developed and calibrated based on present-day relationships, or relationships covering the very recent past (i.e. the past ~100–200 a). Furthermore, many of the approaches have been initially developed for reconstructing climatic changes during the Holocene.

Proxy-based approaches to climatic reconstruction rely on indirect, and in many cases empirical relationships between the proxy indicator and climatic variable of interest. It is therefore clear that they can fail when applied to periods and regions in which climatic conditions differed from climatic states represented in calibration data, or if different combinations of proxy state and climatic conditions occurred in the past than are represented in the modern environment. Centennial- to millennial-scale climatic changes during the last glaciation clearly exceeded the range of climates experienced and recorded in the Alpine region during the past two centuries. As a consequence, transfer functions calibrated in time often have limited use for inferring climatic change >11 ka. Even models which have been calibrated based on a range of varying environmental conditions within the Alpine region using the calibration-in-space approach have to be extended and modified to be applied to records covering the Lateglacial, e.g. by extending calibration datasets to encompass boreal and arctic environments (e.g. Heiri et al., 2011). Uncertainties associated with reconstructions of Glacial and Lateglacial climate based on modern calibration data can be reduced by cross validating reconstructions (e.g. Bradley, 1999). Climatic variations and spatial patterns in climate change that are confirmed by several independent lines of evidence can be interpreted with high confidence. This is in contrast to features that are either documented by single approaches only, or by variations that are not consistently recorded by different proxies. Cross validation is ideally based on several quantitative records, such as the reconstructions of July air temperature inferred from chironomid and pollen assemblages in the same Lateglacial record from Gerzensee, Switzerland (Lotter et al., 2012). Within the prediction errors the two reconstructions agree well in the time window ~15.3–13.8 ka, and absolute estimates of past July temperature in this interval can therefore be interpreted with some confidence. However, reconstructed values deviate in the interval ~13.8–13.0 ka, the chironomids suggesting warmer temperatures than inferred by pollen, pointing towards a potential bias in one or both of these records in this time window. Cross-validation with semi-quantitative or qualitative climate records can also provide independent support for the reliability of climate reconstructions, although the result of this exercise will be less rigorous than an inter-comparison of fully quantitative data (e.g. Heiri et al., 2004).

At present very few systematic cross validation exercises between independent, quantitative approaches are available for the Alpine region for the time window 60–8 ka. The need for inter-calibration and validation also highlights that additional, independent approaches for reconstructing temperature variations 60–8 ka are necessary in order to expand the range of methods available for multi-proxy comparisons. For example, a recently published quantitative temperature reconstruction based on archaeal membrane lipids (TEX86) from Lake Lucerne now provides an additional record for assessing the amplitude of Lateglacial temperature changes in Central Europe (Blaga et al., 2012) that can help to resolve the observed discrepancies between records based on more established approaches.

4.1.5. Records representing variables other than temperature

Our review demonstrates that the vast majority of studies providing quantitative estimates of past climatic change in the study region within the interval 60–8 ka focus on summer temperature (e.g. most biotic proxies, glacier records), or variables believed to mainly reflect past variations in mean annual temperature (e.g. δ18O of precipitation). Very few records are available that are based on established, quantitative approaches and reconstruct other thermal variables (e.g. reflecting autumn temperature: Schmidt et al., 2006) or changes in precipitation and evaporation (e.g. lake level changes: Magny et al., 2003). Again, this problem is not restricted to the Alpine region. However, the Alps and their forelands, with the large number of proxy types recording past temperature change, would be well suited for developing approaches that take advantage of processes influenced by temperature as well as by precipitation and evaporation (e.g. variations in glacier length and volume, pedogenesis, vegetation composition, lake water δ18O) for developing new records of past changes in effective moisture. Again, replicated and cross-validated
temperature reconstructions for different regions of the Alps would be necessary to further develop these approaches. Several recently published records of late and mid Holocene climatic change in the Alpine region (e.g. Kamenik and Schmidt, 2005; Schmidt et al., 2006, 2008) and Late Pleistocene climatic variations in adjacent regions (e.g. Wagner-Cremer and Lotter, 2011) indicate that new approaches to calibrating and interpreting established palaeoenvironmental indicators may also be suited for providing seasonal temperature records from Alpine proxy records for the time window 60–8 ka (e.g. spring temperatures).

4.2. Strategies for the future

Based on the analysis of the present situation, we identify a number of developments that would provide major advancements for quantitative climate reconstruction in the Austrian and Swiss sectors of the Alpine region in the interval 60–8 ka.

(1) Compilations of individual records to continuous reconstructions: Summary datasets describing changes in climatic variables or processes influenced by climate over large sections of the time window 60–8 ka are needed to make the available, fragmentary datasets accessible to the wider user community. Such compilations would not only facilitate the use of Alpine proxy data by palaeoclimatologists interested in climatic change at a European to global scale; they would also highlight where additional records and efforts are necessary to fill gaps in the presently available data. Furthermore, such data compilations would be directly comparable with more continuous Late Quaternary proxy records developed in regions beyond the maximum ice extent of the last glaciation.

(2) Replication and regional reconstructions of past climatic change: Very few late Quaternary climate reconstructions in the Alpine region are based on several, replicated records from the same region (see discussion above). Such records would allow the detection of local artefacts in individual records. Furthermore, replicated records would allow an assessment of the reliability and reproducibility of individual approaches when applied to multiple sites, and the development of new and more robust numerical approaches to develop regional reconstructions (e.g. consensus reconstructions, stacking of records).

(3) Cross-validation of different proxy types and approaches: As outlined above, cross-validation of different approaches provides a rigorous assessment of which aspects of past climate change (e.g., rates of change, amplitudes of change, maximum or minimum values, or direction of change) are supported by different, independent lines of evidence, and which aspects are presently not robustly constrained by multiple approaches. Since all quantitative approaches to reconstruct climatic change may fail if applied to environmental conditions different than those for which an approach has been developed, or if major assumptions behind the approach are violated, such cross-validation exercises are urgently needed. Furthermore, cross-validations may also highlight periods in the past for which different approaches provide diverging results and which may have been characterised by climates different than today. Whether opportunities for cross-validation develop in the near future is obviously related to the quality and reliability of emerging proxy records. Replication and compilation of available records, identified as major aims above, would therefore also greatly facilitate such intercomparisons. Furthermore, it is clear that additional, newly developed proxy indicators and approaches to reconstruction would expand the options for cross-validation. The development of such new approaches should therefore be pursued in the Alpine region.

(4) Reconstruction of past climatic gradients across the Alps: Replicated regional reconstructions of climatic change from several regions would allow an assessment of variations in climatic gradients across the Alpine region. The Alps are an exceptionally interesting area in this context since regional diverging trends of, e.g., temperature or precipitation would allow insights into past circulation patterns and advective effects within the Alpine arch (e.g. Florineth and Schlüchter, 2000). Presently available records are too sparse for most regions, or poorly accessible, complicating such an approach. However, regional differences apparent in some proxy types, e.g. in Younger Dryas ELA between central and outer sectors of the Alps, clearly highlight the potential of such regional, time-transgressive comparisons.

(5) Downscaling of model results to Alpine topography. Climate proxy records play an important role for establishing long-term climate trends and climate linkages, and for understanding leads and lags between climate forcing factors and climatic change. Climate proxy records are also increasingly used for evaluating climate model runs under boundary conditions of the Late Pleistocene or the Holocene (e.g. Renssen et al., 2009). Such comparisons allow an assessment of whether climate models can adequately infer climatic conditions for time periods when climate and climate forcing conditions differed significantly from the modern state. The spatial resolution of most available climate models used to hindcast climate change during the late Quaternary is too coarse for the complex Alpine orography to be adequately represented. Therefore, it is presently difficult to assess whether differences between proxy records from the Alps, influenced by local effects, and climate model hindcasts, typically representing large-scale circulation patterns, are significant. Downscaling of climate model runs 60–8 ka to take into account the Alpine orography would allow a more refined comparison of climate proxy data and model runs. Such runs would highlight under which conditions diverging climatic trends are expected for different sectors of the Alps. Furthermore, they would provide information regarding the climatic sensitivity of different regions and altitudinal belts within the Alps to past and future changes in atmospheric circulation.

(6) Forward modelling of proxies. The majority of approaches to climate reconstruction 60–8 ka rely on empirical relationships between proxy indicators and climatic variables to reconstruct past climatic change. In a second step, climate reconstructions are then compared amongst themselves, with climate model data, or with climate forcing records. An alternative to assess whether a climate scenario is compatible with proxy records is to model the proxy state (e.g. treeline altitude, ELA, permafrost distribution) based on available palaeoclimate data and to compare the result with independent proxy data (e.g. plant macrofossil records obtained from sensitive altitudes, changes in glacier mass or size, evidence for permafrost in LPS). Heiri et al. (2006), Henne et al. (2011) and Schwörer et al. (2013b) used this approach to model expected variations in treeline vegetation in the early and mid Holocene from chironomid-based temperature variations using dynamic vegetation models. The model results were then compared with palaeobotanical proxy data. This forward modelling approach also opens new avenues for assessing whether climate model runs are compatible with proxy records by integrating forward
models with climate modelling (e.g. Schmidt, 2010; Sturm et al., 2010). Furthermore, the approach has the advantage that it can provide insights into past climatic change based on proxy types that are strongly influenced by several climatic variables, as long as independent estimates for the variables other than the one to be reconstructed are available. Replicated proxy compilations for different regions of the Alps, ideally based on multiple, independent approaches, would provide further opportunities to develop forward modelling as a palaeoclimatic tool in the Alpine region. Similarly, downscaled climate model results would provide the opportunity of modelling proxy response and comparing the outcome with observational data. The development of refined modelling techniques for forward modelling landscape processes and ecosystem development would therefore be an important contribution to palaeoenvironmental reconstruction in the Alpine region.

Datasets presented in Figs. 3–5 and 7 (or links to repositories archiving these datasets), together with the pleniglacial dataset shown in Fig. 6 are provided in the Online Supplementary Data.

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Appendix A. Supplementary data

Supplementary data related to this article can be found at http://dx.doi.org/10.1016/j.quascirev.2014.05.021.

References


