The Holocene 15,6 (2005) pp. 789–801

Glacier and lake-level variations in west-central Europe over the last 3500 years

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Received 30 June 2004; revised manuscript accepted 20 November 2004

A HOLOCENE RESEARCH PAPER

Abstract: On the basis of glacier and lake-level records, this paper attempts, for the first time, a comparison between high-resolution palaeohydrological and palaeoglaciological data in west-central Europe over the past 3500 years. A data set of tree-ring width, radiocarbon and archaeological data, in addition to historical sources, were used to reconstruct fluctuations of the Great Aletsch, the Gorner and the Lower Grindelwald glaciers in the Swiss Alps. The three ice-streams experienced nearly synchronous advances at *c.* 1000–600 BC and AD 500–600, 800–900, 1100–1200 and 1300–1860. These glacier fluctuations show strong correspondence with lake-level variations reconstructed in eastern France (Jura mountains and Pre-Alps) and on the Swiss Plateau. This supports the hypothesis of climatically driven fluctuations. Historical data available for the period since AD 1550 reveal, in detail, various meteorological conditions behind the successive glacier advances. However, in agreement with the general trend shown by the historical data, the synchroneity between glacier advances and periods of higher lake level suggests the impact of general winter cooling and an increase in summer moisture as responsible for reinforced feeding of both glaciers and lakes in west-central Europe over the past 3500 years. Finally, a comparison between the Great Aletsch glacier and the residual ^{14}C records supports the hypothesis that variations in solar activity were a major forcing factor of climatic oscillations in west-central Europe during the late Holocene.

Key words: Glacier fluctuations, lake-level variations, historical documents, dendrochronology, radiocarbon dating, climate change, Neoglaciation, 'Little Ice Age', west-central Europe, Alps, Holocene.

Introduction

Most data collected to reconstruct past Holocene climatic oscillations, such as variations in oxygen-isotope ratio (von Grafenstein *et al.*, 1998), tree-line altitude (Bortenschlager, 1970; Tinner *et al.*, 1996; Haas *et al.*, 1998), aquatic plant community (Haas, 1996) or chironomid assemblages (Heiri *et al.*, 2003), preferentially document past changes in temperature. Pollen data offer the advantage of giving information on both temperature and precipitation, but problems arise when reconstructing past precipitation from modern analogue methods because, in Europe, moisture in the growing season is rarely the main limiting factor for vegetation on a regional scale, and modern analogues tend to span a wide range of precipitation estimates (Guiot *et al.*, 1993). However, from the perspective of a global warming scenario (Vörösmarty *et al.*, 2000), a crucial question in palaeoclimatology is the variations in the hydrological cycle associated with climatic changes.

As a contribution to fill this gap, this paper presents high-resolution records of variations in glacier size in the Swiss Alps and lake-level fluctuations in the Jura mountains, the northern French Pre-Alps and the Swiss Plateau over the last 3500 years. These two distinct data sets provide an opportunity to check changes in the thermal and hydrological cycles associated with major climatic oscillations previously recognized for this time window, i.e., the cooling at the sub-Boreal–sub-Atlantic transition (van Geel *et al.*, 1996), the Roman and Mediaeval warm periods (Holzhauser, 1997; Holzhauser and Zumbühl, 2003) and the 'Little Ice Age' (LIA, see Grove, 1988; Holzhauser and Zumbühl, 1999). They also offer the

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possibility to test hypotheses about the driving factors behind climate changes during this period, while instrumental and historical data over the last centuries allow progress in our understanding of weather conditions associated with hydrological changes.

Glacier fluctuations

Investigations of Holocene glacier fluctuations in the Swiss Alps have been carried out for a number of years (e.g., Röthlisberger, 1976; Zumbühl, 1980; Zumbühl *et al*., 1983; Holzhauser, 1984; Zumbühl and Holzhauser, 1988; Furrer, 1991; Holzhauser and Zumbühl, 1996, 2003). Because glaciers are sensitive to climatic variations (Intergovernmental Panel on Climate Change (IPCC), 2001), the reconstruction of pre-industrial glacier fluctuations therefore reveals the natural range of Holocene climate variability against which the present-day climatic situation can be judged. The glacier records also provide a very useful base for glacier modelling as well as calibration and cross-validations for glacier–climate models (Haeberli and Holzhauser, 2003).

Recent studies on the Great Aletsch and Gorner glaciers in the Alps of Valais, as well as on the Lower Grindelwald glacier in the Bernese Alps (Figure 1), have shed light on glacier fluctuations during the last 3500 years (Figure 2). These glaciers belong to the most famous and most investigated ice-streams in the Swiss Alps (Zumbühl, 1980; Holzhauser, 1984, 2001). They have been the subject of scientific as well as artistic interest for many centuries. At their maximal extensions, they penetrated below the timberline and have even occasionally reached inhabited areas, resulting in massive destruction. Extensive historical records, loss of buildings, woods and pasture form the basis for the historical methods used for the reconstruction of glacier length fluctuations through time.

Methods

The methods used for reconstruction differ in time resolution (yearly to century-scale) and time windows that they document.

Glaciological methods provide direct observations and measurements of volume, surface area and changes in tongue length over the last 120 years (Haeberli, 1995; Müller-Lemans *et al*., 1996; Hoelzle and Trindler, 1998).

Historical and archaeological methods involve the collection of data from written and pictorial historical records, which may include paintings, sketches, engravings and photographs as well as written accounts (e.g., chronicles and scientific works about glaciers and the Alps) and topographical maps. These records give a very detailed picture of glacial fluctuations in the Swiss Alps going back at least 450 years (Zumbühl, 1980; Holzhauser and Zumbühl, 2002). The historical method can achieve a resolution of decades or, in some cases, even single years. The archaeological method is based on (1) finding human traces such as foundations and wooden beams from buildings, old alpine routes or remains of abandoned irrigation channels, and (2) connecting them with glacier history. This method can, in certain areas, provide evidence of glacial fluctuations as far back as the late Middle Ages (Röthlisberger, 1976; Holzhauser, 1984).

The glacio-morphological method entails the search for dateable organic material such as soils and crushed trees within and at the edge of glacier forefields. This material is only revealed once the ice has retreated. The dating of fossil soils (palaeosols) and wood makes it possible to establish a chronology of glacier fluctuations throughout the entire Holocene (Röthlisberger, 1976; Schneebeli, 1976; Holzhauser, 1984; Nicolussi and Patzelt, 1996, 2000; Hormes *et al*., 2001). The interpretation of radiocarbon dates of fossil soils is not so straightforward because soil formed at the beginning of a development period can often be mixed with soil formed later in the period of glacier contraction. This can result in the measured age lying somewhere between these limits (Gamper, 1985; Geyh, 1986).

The dating of fossil wood (tree trunks and stumps, roots and macro-remains) is mainly achieved with the radiocarbon method. The dating of the outer tree-rings from trees filled by glacier advances gives an approximate time of death of the tree and therefore the point in time of glacier expansion. Only roots, stumps and trunks of trees found in their original position (*in situ*) are suitable for the exact reconstruction of changes in glacier size (Holzhauser, 1997, 2002). The total number of tree-rings gives a minimum time in which the tree

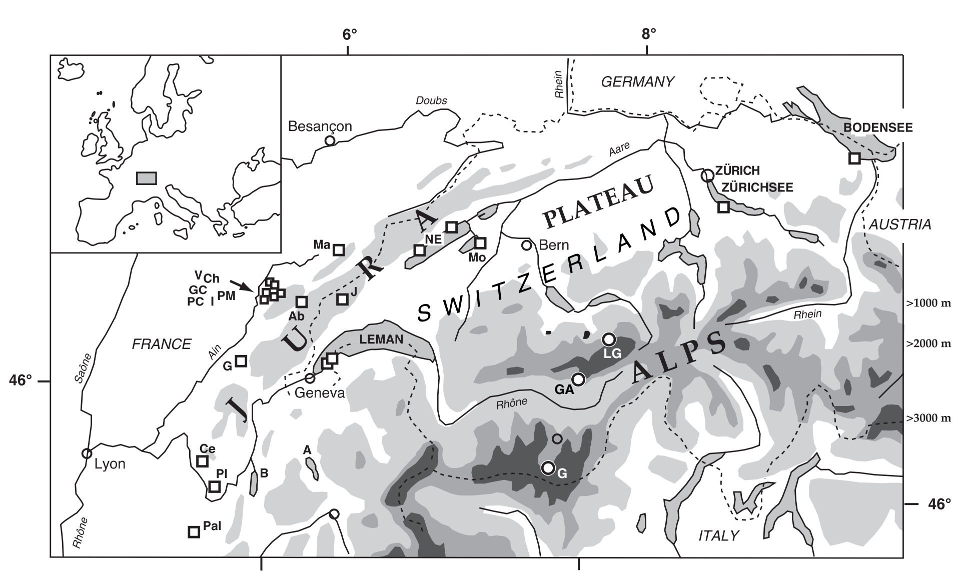

Figure 1 Geographical location of study sites. Glaciers: GA, Great Aletsch glacier; G, Gorner glacier; LG, Lower Grindelwald glacier. Lakes: A, lake Annecy; Ab, lake Abbaye; B, lake Bourget; Ce, lake Cerin; Ch, lake Chambly; GC, lake Grand Clairvaux; Ge, lake Genin; I, lake Ilay; J, lake Joux; Ma, lake Malpas; Mo, lake Morat; NE, lake Neuchâtel; Pal, lake Paladru; PC, lake Petit Clairvaux; PM, lake Petit Maclu; Pl, lake Pluvis; V, lake Val

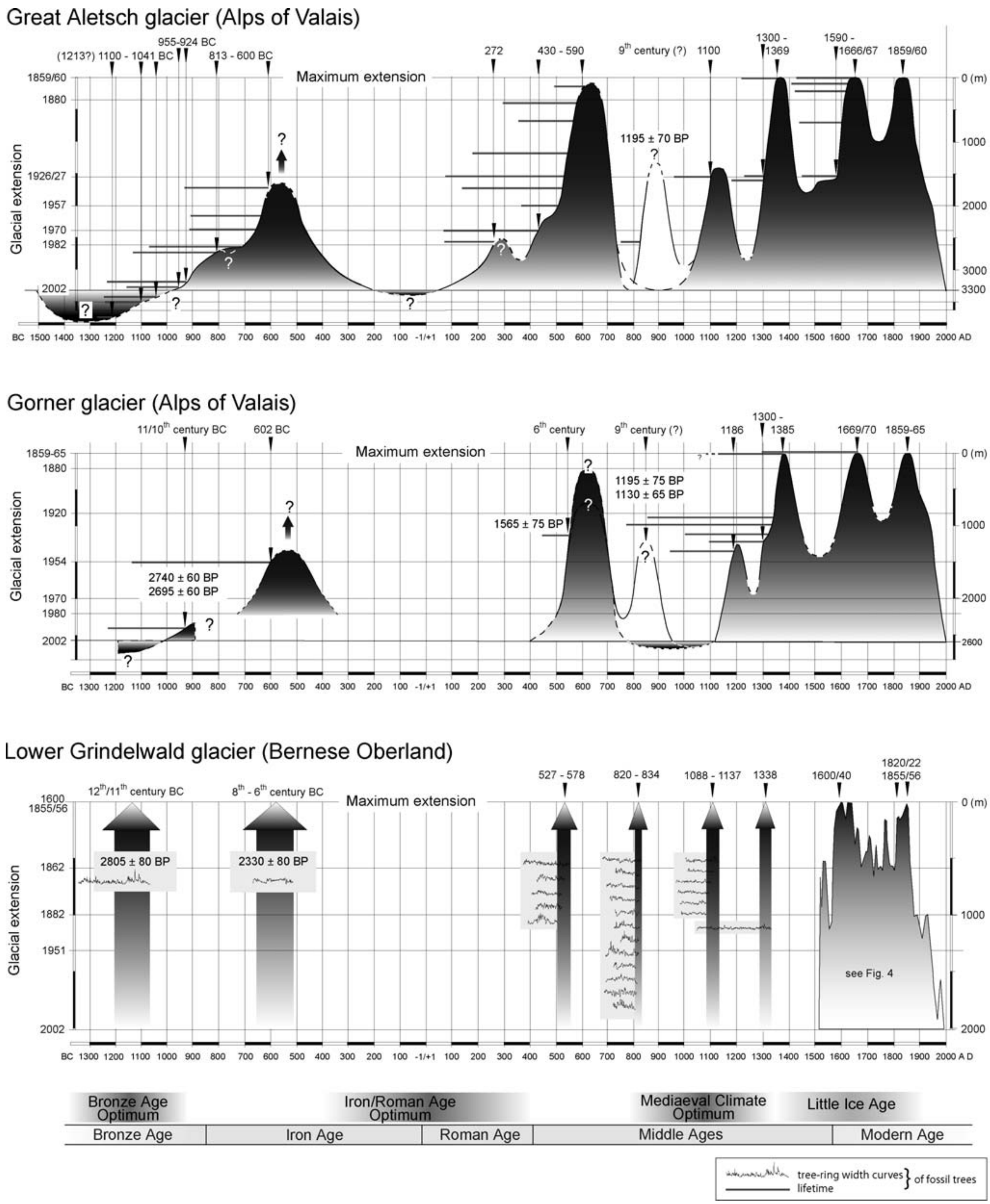

Figure 2 Fluctuations of the Great Aletsch, the Gorner and the Lower Grindelwald glaciers over the last 3500 years

site was not covered by a glacier. It is thus possible to find the length of time between two glacial advances with relative accuracy. A sample found close to the actual glacier edge would prove that the glacier, at some time in the past, was shorter than it is today.

Radiocarbon dating of fossil soils and trees has, at best, an accuracy of 100–200 years. If a well-preserved wood sample possesses enough tree-rings, it is possible to date the lifetime of the tree with dendrochronology. Dendrochronological analysis of fossil trees has supplied much reliable evidence of glacier fluctuations through its proven precision (resolution of a single year), impossible to obtain with the radiocarbon method alone. A precondition to the exact dating of trees from the present time through to the Middle Ages is an absolute standardized tree-ring width chronology. Such an absolute chronology, (MKWALLIS) covering the years 2879 BC to AD 2000, was established with tree-ring width curves from living larch trees (*Larix decidua* Mill.) located at the Alpine timberline in the Swiss Alps (Holzhauser, 1997 and Holzhauser, unpublished data).

Sensitivity to climate changes and climatic significance

Advance and retreat of mountain glaciers during historical and Holocene time periods reflect pre-industrial changes in ice mass and corresponding energy fluxes between the atmosphere and the earth (Haeberli and Holzhauser, 2003). The behaviour of glaciers is highly dependent upon climatic conditions and they react according to the sum of complex interactions of climatic parameters (e.g., temperature, precipitation, solar radiation) and topographical conditions (Hoinkes and Steinacker, 1975; Kuhn, 1980, 1990; Haeberli, 1994; Oerlemans, 2001; Hoelzle *et al.*, 2003). Either they increase in length or, as under present circumstances, they melt and shorten. On the one hand, glaciers shorten in length if temperatures rise or precipitation decreases or, on the other hand, they advance if temperatures decrease and/or precipitation increases (Winkler, 2002; Nesje and Dahl, 2003).

The resulting change in glacier length (advance or retreat) is via mass balance, an indirect, filtered and delayed signal of these climatic fluctuations. Basically, the reaction and response time of a glacier is proportional to its size. After a certain reaction time, in the case of Alpine glaciers ranging from a few years to several decades, glacier length changes, finally reaching a new equilibrium after a response time from several years to about 100 years (Haeberli, 1994, 1995; Haeberli and Hoelzle, 1995). Small and medium-sized glaciers, e.g., the Lower Grindelwald glacier, may show changes within a few years and are therefore sensitive to short-term climatic changes (Zumbühl *et al.*, 1983). Larger glaciers such as the Great Aletsch glacier need much more time to react and indicate long-term climatic fluctuations. It has been calculated that the Great Aletsch glacier has a reaction time of about 24 years and the Gorner glacier 19 years (Müller, 1988). The response time is much longer. In the case of the Great Aletsch glacier, it is suggested that it is an order of 50 to 100 years (Haeberli and Holzhauser, 2003). The present-day position of the glacier front is therefore a reflection of the climatic conditions of past decades.

In most historical works concerning glacier fluctuations only maximum extensions are noted. Such an approach indicates very little about the magnitude of the corresponding climatic change and the resulting mass change of the glaciers. Thus, it is necessary to reconstruct both the minimum and maximum positions of the glacier terminal-point at the beginning and end of each individual glacier-advance period. Only with this information can the total amplitude of glacier movement and the associated mass and climatic change be measured (Haeberli and Holzhauser, 2003).

Results

Fluctuations of the Great Aletsch glacier (Alps of Valais)

The Great Aletsch glacier with a length of around 23 km and surface area of 81.69 km^2 in 1998 (Paul, 2003) is the largest glacier in the European Alps. Since reaching a maximum length in 1856/60, the glacier has receded by 3.4 km, representing around 24 m/yr (Figure 3). The tongue of the Great Aletsch glacier has, in the past, reached deep into the pine forest, sometimes coming very close to human-inhabitated areas. During recent maximums, mountain huts have been destroyed, an irrigation channel has been made unusable and pine forest and arable land have been covered.

Using all the methods mentioned above, it has been possible to reconstruct the changes in length of the glacier tongue over the past 3500 years. Reconstruction is based on absolute (calendar year) dating. The main method used was the dating of fossil trees found within the forefield, some of them actually still *in situ*. The most recent segment of the curve presented in Figure 2, from the twelfth century AD on, is based on dendrochronological and archaeological evidence as well as visual and written historical sources. The period from the twelfth century AD backwards is reconstructed only by means of dendrochronologically absolute dated fossil wood.

From the late Bronze Age to the Middle Ages, evidence from dendrochronologically absolute dated trees is obtained not only of growth phases of glaciers, some quite marked (*c*. 813–600 BC), but also of periods when glacier size was similar to or smaller than today. During the late Bronze Age Optimum from 1350 to 1250 BC, the Great Aletsch glacier was approximately 1000 m shorter than it is today. The period from 1450 to 1250 BC has been recogized as a warm-dry phase in other Alpine and Northern Hemisphere proxies (Tinner *et al.*, 2003). During the Iron/Roman Age Optimum between *c*. 200 BC and AD 50, the glacier reached today's extent or was even somewhat shorter than today. In the early Middle Ages (around AD 750), the Great Aletsch glacier reached its present extent.

The Mediaeval Warm Period, from around AD 800 to the onset of the LIA around AD 1300, was interrupted by two weak advances in the ninth (not certain because based only on radiocarbon dating) and the twelfth centuries AD (around AD 1100). Precise dendrochronological dating shows a powerful advance at the beginning of the LIA, around AD 1300. The LIA is characterized by three successive peaks: a first maximum after 1369 (in the 1370s), a second betweeen 1670 and 1680, and a third at 1859/60.

Sparse evidence is available from the fifteenth century AD. Written documents indicate that the glacier was of a size similar to that of the 1930s. A very small advance around 1500 has been dated by both dendrochronology and archaeology (destruction of an irrigation channel). The dendrochronological analysis of *in situ* wood samples found in the glacier forefield has made possible a very exact reconstruction of the advance that occurred in the last third of the sixteenth century up until the maximum in 1670–1680 (Holzhauser and Zumbühl, 1999; Holzhauser, 2002). There are some historical pictures and texts from the eighteenth and nineteenth centuries that, more or less, document the fluctuations of the Great Aletsch glacier in this period. The tongue-length has been measured yearly since 1892, making the last century extremely well documented (see annual reports of the Swiss glaciers in Die Alpen, Zeitschift des Schweizer Alpen-Clubs, Stämpfli AG, Bern).

At Great Aletsch glacier effects of high-frequency (interannual and decadal) climate variability are – contrary to the case of the Lower Grindelwald glacier – filtered out by the slow dynamic response of the relatively long glacier. Such filtering results in a strong smoothing of the signal and an extremely clear reflection of secular trends (Haeberli and Holzhauser, 2003). The glacier tongue is now approaching the previous minimum recorded length, as documented between 200 BC and AD 50. In this context, it must be borne in mind that the dynamics of the Great Aletsch glacier tongue not only constitute a smoothed but also a slightly delayed function of direct climate and mass balance forcing. In this case, the corresponding time lag is estimated at a few decades, which means that the glacier tongue would have to be hundreds of metres shorter than now in order to adjust to conditions of the year 2004. In view of the rapid warming during the past two decades, it is highly probable that, in the near future, the previous minimum extent of late Bronze Age may be reached or even markedly exceeded.

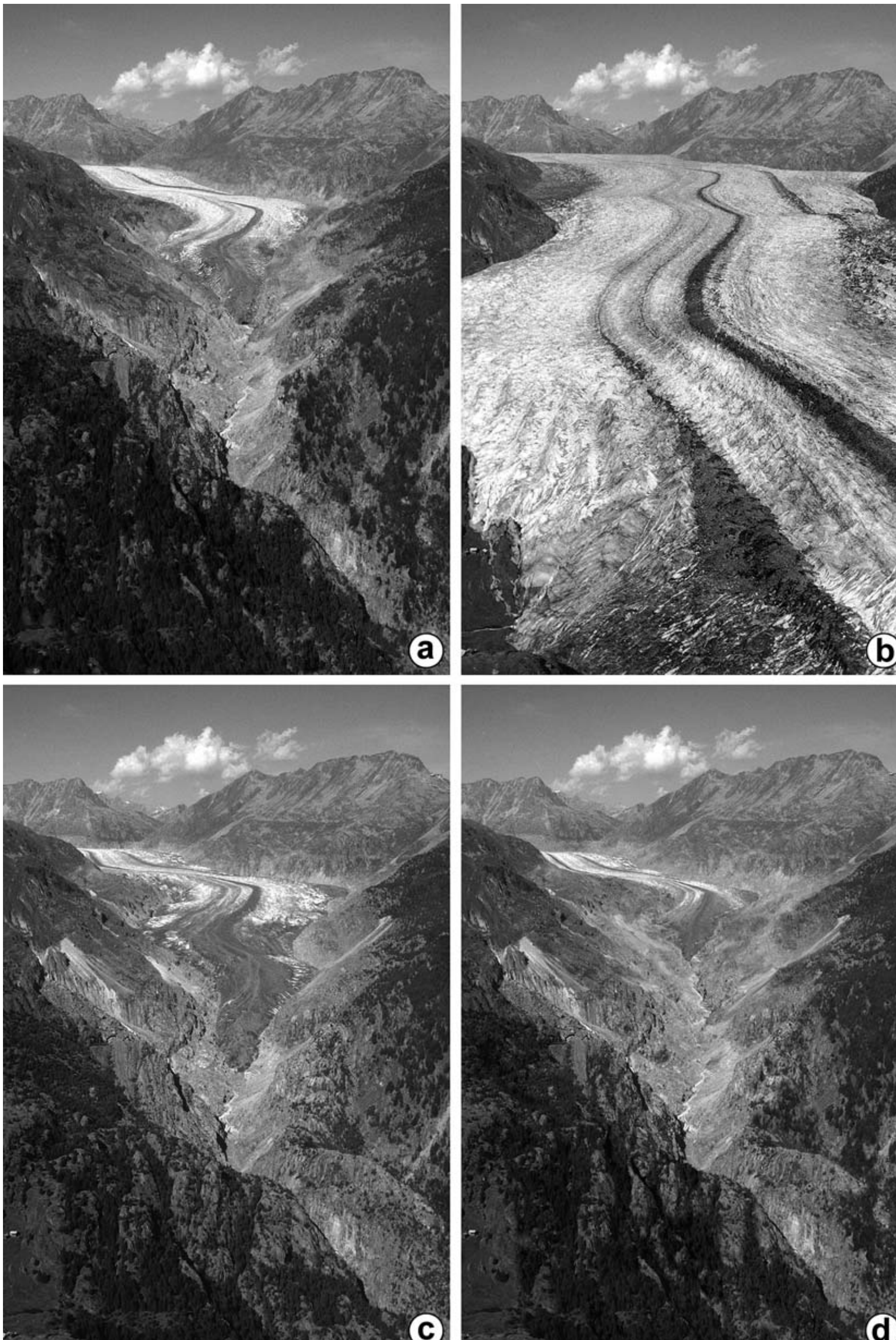

Figure 3 The Great Aletsch glacier within four periods: (a) Iron/Roman Age Optimum (c. 200 BC–AD 50; (b) AD 1856; (c) AD 2000 and (d) around AD 2050 (comparable with the Bronze Age Optimum). The glacier extensions in (a) and (d) are reconstructed on the basis of glaciological evidence. Original photograph from 1856: Frédéric Martens, Alpine Club Library London. Photomontage: H. Holzhauser

Fluctuations of the Gorner glacier (Alps of Valais)

The Gorner glacier is the second largest glacier in the Swiss Alps (surface 57 km^2 in 1998; Paul, 2003, see also Maisch *et al.*, 2000). During its maximum extension the name 'Bodengletscher' was given to the lowest section, which was easily visible from Zermatt and extended over arable land. Thanks to high quality written and visual records (Holzhauser, 2001), and in particular radiocarbon and dendrochronologically absolute dated fossil trees from the forefield, it has been possible to reconstruct a nearly complete and accurate picture of changes in the length of the Gorner glacier over the past 3100 years (Figure 2).

The two advances in the Bronze Age (tenth century BC) and in the early Middle Ages (sixth century AD) are radiocarbon dated. The dendrochronologically dated advance around 602 BC corresponds closely to the advance of the Great Aletsch glacier at this time. The advance in the ninth century AD is radiocarbon dated and is therefore uncertain, as in the case of the Great Aletsch glacier curve.

From the (uncertain) advance in the ninth century AD up until the beginning of the LIA around AD 1300, the end of the tongue was apparently not extended as far as it was in 1940. During the Mediaeval Climate Optimum, the glacier advanced weakly in the twelfth century AD and in AD 1186 reached an extent comparable with that of *c*. 1950.

At the onset of the LIA, the Gorner glacier advanced relatively slowly between 1327 and 1341 by 9 m/yr approximately. From 1341 until it reached its first maximum length within the LIA in 1385, it began to move at a faster rate of around 20 m/yr. The tongue thus advanced at an average of 17 m/yr between 1327 and 1385, making a total of 1000 m over that period. At no other glacier in the Swiss Alps are the fourteenth-century advance and the Mediaeval Climatic Optimum so well documented as at the Gorner glacier. After a period of retreat, the Gorner glacier reached its second maximum extent during the LIA, in 1669–70. From the end of the eighteenth century onwards, the history of the Gorner glacier is well documented with historical data. Around 1791, the Gorner glacier was approximately the same length as it was at around 1900. At the beginning of the nineteenth century, the glacier started a long-term and more or less continuous phase of growth that lasted until 1859, when the glacier reached the third LIA maximum. During this period, the tongue advanced at an average of 10 m/yr, destroying many houses and farm buildings as well as valuable farmland in the process (Holzhauser, 2001). Since 1865, the tongue has receded 2600 m, and the 'Bodengletscher' has completely disappeared.

Fluctuations of the Lower Grindelwald glacier (Bernese Alps)

Owing to the fact that during the LIA the tongue of the Lower Grindelwald glacier (length today less than 9 km; surface area 18.7 km^2 in 1998; Paul, 2003) often extended as far as the valley floor and was thus in close proximity to the village, it has been one of the best known and most visited glaciers in the whole of the Alps since the eighteenth century. Owing to the mass of information sources covering the last 450 years, a degree of precision can be achieved that is simply not possible in the case of most of the other glaciers in the world. Equally outstanding is the more than 360 paintings, drawings and photographs up to 1900 and often also the qua1ity of the historical written accounts and visual records of the glacier (Zumbühl, 1980; Zumbühl *et al*., 1983). Fossil soils and trees found particularly in the lateral moraines at Stieregg and Zäsenberg (Holzhauser and Zumbühl, 1996) complete this evidence and enable the reconstruction of the glacier history over the past 3100 years.

It has been possible to identify a total of six marked growth periods for the Lower Grindelwald glacier between the end of the Bronze Age and the high Middle Ages (Holzhauser and Zumbühl, 1996). The two periods of advance in the eleventh and in the seventh/sixth centuries BC were fixed only by radiocarbon dating and seem to correspond to periods of advance of the Great Aletsch glacier. From the sixth century AD to the fourteenth century AD, the advance periods of the Lower Grindelwald glacier are absolutely dated by means of dendrochronology (Figure 2). The late Middle Ages growth phase around 1338 belongs to the three peaks of the LIA (*c*. 1300–1850).

It is not possible to draw a continuous and complete curve of the fluctuations of the Lower Grindelwald glacier. Unfortunately the geometry (minimum and maximum extension at the beginning and end, respectively, of the different growing periods) of this glacier during the various periods of advance from 1100 BC to the fourteenth century AD is not known.

The development of the glacier after 1500 can be summarized as follows (Figure 4). Around 1540 there was only a small advance. A long-term growth period began around 1575 and lasted approximately until 1600. This activity led to a glacier advance of about 1000 m, so the ice stream reached its maximum extent during the LIA. From the seventeenth to the middle of the nineteenth century, a relatively long period, the glacier remained stable, extending as far as the Schopf rock terraces around 1250 m below the present tongue. In the same period there were at least six relatively short-term advances of between 400 and 600 m, which led to the formation of a marked fan-shaped tongue ('Schweif') which extended far down into the valley. Three maximum lengths were reached (in 1778–79, 1820–22 and 1855–56) and during three periods the 'Schweif' was shorter, e.g., 1669 (see illustrations in Holzhauser and Zumbühl, 2003). Since the end of the LIA the Lower Grindelwald glacier has receded by about 2 km to its present limit well up in the gorge.

Lake-level fluctuations

Lake levels are influenced by climate variables affecting evaporation as well as precipitation but can also be modified by various local, non-climatic factors such as diversion of lake tributaries or anthropogenic forest clearance in the catchment area. However, synchronous changes in lake levels within a region can be assumed to be climatically driven (Digerfeldt, 1988; Magny, 1998). Thus, lake-level records have been used in Europe to establish spatial patterns of Late Quaternary changes in moisture balance and atmospheric circulation (Harrison and Digerfeldt, 1993), or to refine quantification of climatic parameters based on pollen data (Guiot *et al*., 1993; Magny *et al*., 2001, 2003). Over the last 20 years, systematic sedimentological investigations on Holocene lake-level fluctuations have been carried out in a region composed of the Jura mountains, the northern French Pre-Alps and the Swiss Plateau (Figure 1).

Methods

The reconstruction of past changes in lake level presented in Figure 5 are based on a specific sedimentological method extensively described elsewhere (Magny, 1992, 1998). This method relies on (1) a modern analogue approach in which the sediment record is interpreted by comparison with the types of deposits accumulating today in the Jurassian and subalpine lakes, and (2) a combination of several parameters such as the texture, the lithology and the frequency of different carbonate concretion morphotypes of biochemical origin. Each morphotype shows a specific spatial distribution in relation to the increase in water depth and changes in aquatic vegetation belts from the shore to the extremity of the littoral platform, with successive domination of oncolites (nearshore areas with shallow water and high-energy environment), cauliflower-like forms (littoral platform), plate-like concretions (encrustations of leaves from the Potamogetonion and Nymphaeion belts) and finally tube-like concretions (stem encrustations from the Characeae belt on the platform slope) (Magny, 1992, 1998). In addition to these sediment markers, the geometry of the sediment layers can highlight erosion

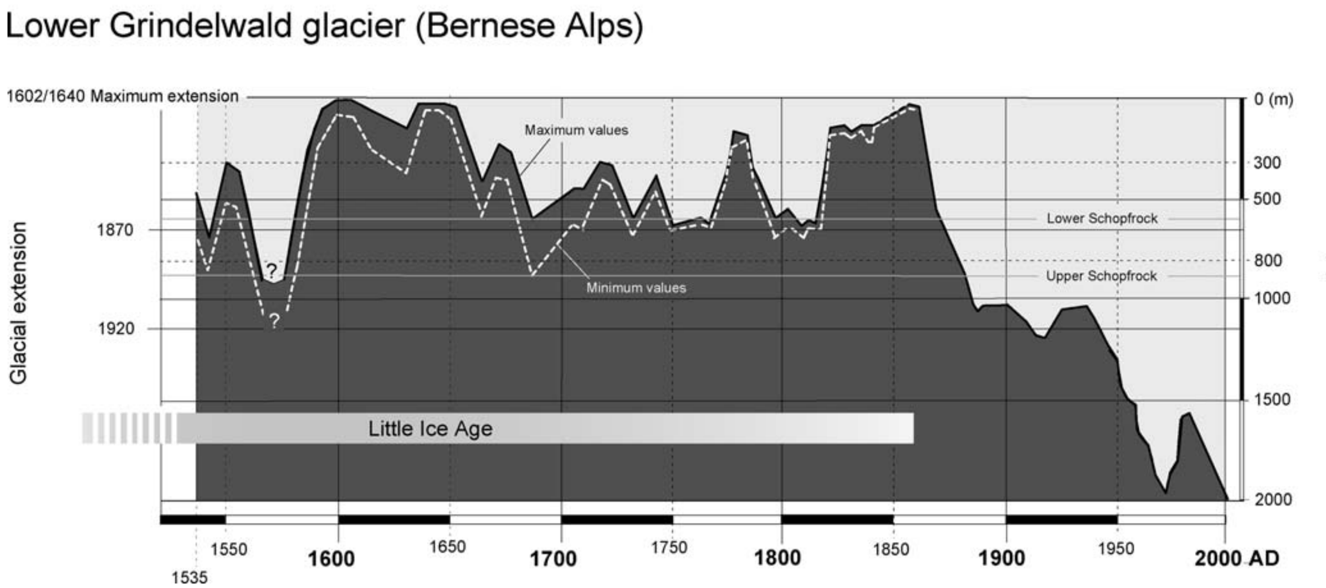

Figure 4 Fluctuations of the Lower Grindelwald glacier during the modern age, reconstructed with the help of historical documents (after Zumbühl *et al.*, 1983; Holzhauser and Zumbühl, 2003)

surfaces resulting from a lowering of lake-level and sediment limit (Digerfeldt, 1988). A contiguous high-resolution subsampling of sediment sequences made the identification of short-lived events possible.

Three methods were used for absolute dating of the lake-level changes as follows:

(1) radiocarbon dating from peat, detritus peat and terrestrial plant macrofossils;
(2) tree-ring dating from wooden posts used by prehistoric and historical people for dwelling construction and well preserved in lake-shore archaeological sites (Magny, 2004; Magny *et al.*, 1995). Rests of these posts are interbedded into archaeological layers or, if preserved in vertical position, they are correlated with archaeological layers on the basis of stratigraphic features;
(3) archaeological dating from human artefacts such as pieces of Roman amphoras found on archaeological sites (Magny and Richard, 1985).

Sensitivity to climate changes and climatic significance

Most studied lakes are small, characterized by a common nivo-pluvial regime with lake-level maximums during rainy seasons and at snow melting (spring and autumn) and minimums during the summer. They show a great sensitivity to variations in water supply as shown by seasonal changes in water level and can be considered to react quasi-immediately to climatic events.

Lake-level fluctuations can be assumed to reflect variations in the difference between precipitation and evapotranspiration, but closed basins are rare in temperate zones and all the lake-level data used here were obtained from lakes with inflow and outlets. Hence, it is difficult to reconstruct temperature and precipitation from lake-level fluctuations alone, because a positive water balance may reflect a decrease in precipitation and/or in temperature. Moreover, other factors such as cloudiness and wind have a direct impact on lake evaporation (Hostetler and Benson, 1990). However, pronounced falls in the level of lakes without glacial water supply in the Jura and on the Swiss Plateau today are often induced by prolonged warm and dry summers. Thus, as a first approximation, this observation suggests that lake-level lowering may reflect a negative summer water balance resulting from a decrease in water supply and a longer summer season with stronger evaporation and evapotranspiration (Magny, 1993c). This is supported by quantitative reconstruction of climatic parameters using both pollen and lake-level data for the first half of the Holocene (Magny *et al.*, 2001, 2003).

Results

Figure 5 presents a synthetic regional pattern of late Holocene lake-level fluctuations established from data collected in 20 lakes. On the basis of a combination of 71 radiocarbon, tree-ring and archaeological dates, the age uncertainty in the chronology of this lake-level record can be assessed to ± 50 years (Magny, 2004). It gives evidence for six successive phases of higher lake level as follows.

(1) The onset of phase 6 of higher lake level, which developed between 1550 and 1150 BC, coincided with the cessation of Early Bronze Age lake-shore villages around the Jurassian and Subalpine lakes (Magny, 1993b). It preceded a lowering phase interrupted by two short-lived episodes of rising levels well dated by tree-ring study to 1050–1000 and 950–900 BC.
(2) Phase 5 (800–400 BC) began immediately after the abandonment of the Late Bronze Age lake-dwellings at 815 BC (Marguet and Billaud, 1997). It corresponds to a widespread climate reversal identified at the sub-Boreal–sub-Atlantic transition (van Geel *et al.*, 1996). Several records indicate a bipartition (Magny and Richoz, 1998).
(3) Phase 4 marks a rise in lake level peaking at *c.* AD 150–250 and has an equivalent in a phase of stronger discharge in the middle and lower Rhone valley (Bruneton *et al.*, 2001; Berger, 2003).
(4) Dated to *c.* AD 650–850, phase 3 is composed of two successive high lake-level episodes as recorded in Lake Petit Clairvaux (Magny, 2004).

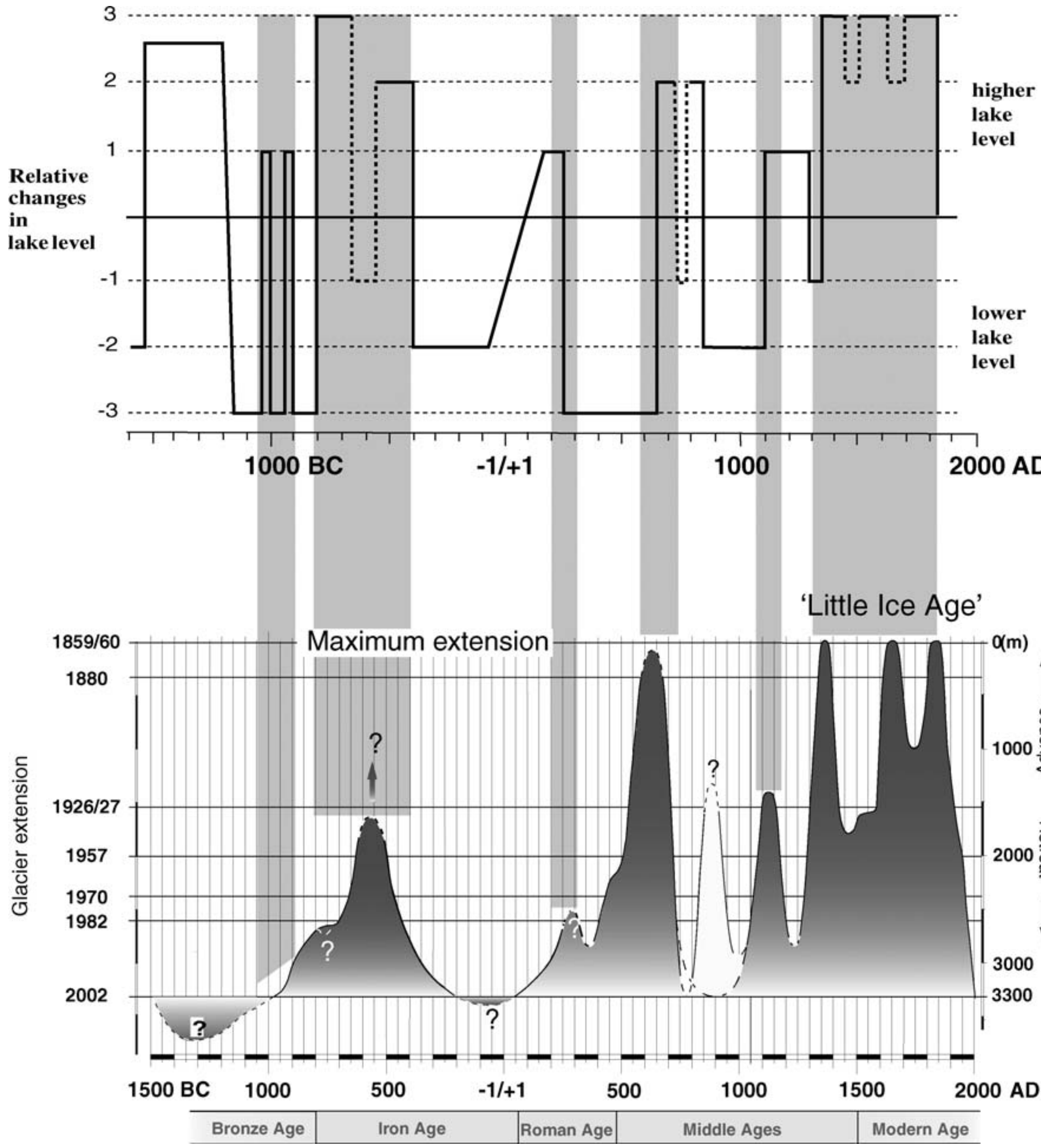

Figure 5 Comparison between the Great Aletsch glacier (bottom; Switzerland) and the west-central European lake-level records (top; after Magny, 2004) over the last 3500 years. Dotted lines in the lake-level record indicate an uncertain chronology

(5) Dated to *c*. AD 1200–1300, phase 2 could have begun as early as AD 1100 as shown by regional lake-level data (Magny, 2004).
(6) Phase 1 began a little before AD 1394 (tree-ring date) as documented by a sediment sequence of Lake Ilay, Jura (Magny, 2004). This last phase marks a long-lived phase of higher lake-levels coinciding with the LIA. Some records indicate a tripartition of this phase which preceded a subcontemporaneous lowering. Based on instrumental data, the historical lake-level record of Lake Neuchâtel clearly shows that before the artificial correction in 1876–1879, the water level experienced a mean lowering by *c*. 0.7 m during the period 1853–1876 in comparison with the relatively higher water level during the period 1817–1853, i.e., during the final part of the LIA (Quartier, 1948).

Thus, the Late Holocene appears to be punctuated by two major phases of higher lake level at 1550–1150 and 800–400 BC (episodes 5 and 1) and two periods of pronounced lowering at 1150–800 and 250–650 BC.

Discussion

Comparison of glacier and lake-level records

Figure 2 shows evidence that, despite differences in size and location, the variations of the Great Aletsch, the Gorner and the Lower Grindelwald glaciers show strong similarities over the last 3500 years. As far as can be judged from available data, the three ice-streams experienced nearly synchronous advances at *c*. 1000–600 BC, AD 500–600, 800–900, 1100–1200 and 1300–1860. The lack of data documenting any advance of the Lower Grindelwald and Gorner glaciers from 400 BC to AD 400 may be paralleled with the long recession phase of the Great Aletsch glacier at this time, and hence interpreted as a possible indication that neither glacier experienced any expansion during this period. Moreover, the advance of the Lower Grindelwald glacier, well dated by tree rings to AD 820–834, supports the reality of expansions of the Great Aletsch and the Gorner glaciers during the ninth century AD, which were dated only by radiocarbon. In addition, the three glacier records show a similar pattern with major, longer-lived advances peaking at *c.* 600 BC, AD 590–600 and between AD 1300 and 1860. The last one coincided with the LIA and is composed of three distinct maximums, while the small differences in the chronology can be attributed to differences in response time.

The comparison of the Great Aletsch glacier record with the lake-level record reconstructed for the Swiss Plateau, the Jura mountains and the northern French Pre-Alps highlights strong correspondences between both the palaeohydrological and palaeoglaciological records (Figure 5). In general, the timing of the successive events and their respective magnitude most often show quite similar patterns. Glacier advances coincided with phases of higher lake level. Only two discrepancies arise (1) in the absence of a lake-level maximum during the ninth century AD in contrast with the glacier expansion at this time, and (2) in the lake-level minimum at AD 250–650 against an earlier glacier minimum culminating at *c.* 200 BC–AD 50. Apart from these differencies, both the records show the same tripartition of the LIA maximum while the bipartition of the lake-level maximum at 800–400 BC may have an equivalent in the two successive advances of the Great Aletsch glacier at 813 and 600 BC. Minor divergences in the timing of events documented by lakes and glaciers may reflect differences in response time to climate factors and age uncertainty in the lake-level record.

Significance for possible forcing factors of the late-Holocene climatic oscillations

While strong similarities between the palaeohydrological and palaeoglaciological records as evidenced in Figure 5 support the hypothesis of events driven by over-regional climatic changes, another topic of discussion is the possible forcing factors of these climatic oscillations. Figure 6 presents a comparison of the Great Aletsch glacier and the atmospheric ^{14}C residual series based on tree-ring records (Stuiver *et al.*, 1998), which may be considered as a proxy record for past changes in solar activity (Hoyt and Schatten, 1997) governing 80 and 200 years periodicities, and probably a 2400-yr quasi-cycle (Renssen *et al.*, 2000; Vasiliev and Dergachev, 2002). However, variations in the solar irradiance are not the only causes of changes in the residual ^{14}C record, which also reflects exchanges with other ^{14}C reservoirs, i.e., the ocean and the biosphere, possibly in relation to climate oscillations. But, variations in ^{10}Be appear to be similar in shape to those of ^{14}C; these similarities suggest that the variations in both the isotopes are primarily driven by changes in production and reflect a strong influence of variations in solar activity (Muscheler, 1999, 2000; Beer, 2000; Renssen *et al.*, 2001).

In general, correlations appear between these records: glacier maximums coincided with radiocarbon peaks, i.e., periods of weaker solar activity. This suggests a possible solar origin of the climate oscillations punctuating the last 3500 years in west-central Europe, in agreement with previous studies (Denton and Karlén, 1973; Magny, 1993a; van Geel *et al.*, 1996; Bond *et al.*, 2001). Certain delays of the glacier maximums in comparison with the radiocarbon peaks, for instance the glacier maximums dated to AD 1100–1200 and possibly related to the ^{14}C peak at AD 1050, may reflect an age uncertainty in the palaeoenvironmental records or a lag due to a response time of the glacier to climatic change. Problems with the chronology and gaps in the documentation may also explain some discrepancies such as the absence of glacier advance when a major radiocarbon peak developed at *c.* 350 BC. Moreover, unless a response time for the Great Aletsch glacier can be assumed to be irregular in time and also greater than 150 years, no direct correlation appears during the LIA between the three ^{14}C peaks at AD 1300, 1500 and 1700 and the three glacier maximums well dated by tree-ring and historical data to around AD 1380, 1666–1670 (1600–1640 in the case of the Lower Grindelwald glacier) and 1855–1860. All major and minor discrepancies may also reflect the possible interference of other climatic forcing factors or mechanisms superimposed on a general solar forcing. Thus, explosive volcanism can cause short-lived anomalous low summer temperatures (e.g., the Tambora eruption in 1816).

Synoptic climatology and climatic implications of the late-Holocene glacier and lake-level records

A last but complex topic of discussion is the significance of the past glacier and lake-level fluctuations, presented in this study, in terms of weather conditions and regimes over west-central Europe. The difficulty in using glaciers as climatic indicators lies in the complex chain of processes (energy and mass balance changes, geometry, flows, length changes) linking glacier reaction to climatic change (Haeberli, 1994). Glaciers in mountain areas are highly sensitive to climate changes and thus provide one of nature's clearest signals of warming or cooling and/or dry and wet climate periods. Thus, in a simplified manner, one can say that the quasi periodical fluctuations of Alpine glaciers were driven by glacier-hostile (warm/dry) and glacier-friendly (cool/wet) periods.

Yet, a detailed examination of historical data suggests a rather more complex situation. As Wanner *et al.* (2000) demonstrated, the principal mechanism causing glacier advance periods (Little Ice Age Type Events, LIATES) are driven by different forcing characteristics showing specific and varying temperature and precipitation courses during winter and summer. For example, the distinct advance of Alpine glaciers at the end of the sixteenth century leading to the second maximum extension within the LIA was a result of predominant low winter and summer temperatures and summer snow falls in the higher Alps (Pfister, 1984). The period of the last LIA advance in the middle of the nineteenth century was also characterized in fact by low temperatures, especially during winter, and a peak in humidity around the 1840s that brought enormous amounts of winter precipitation with snow (Wanner *et al.*, 2000). It can also be hypothesized that the mid-nineteenth century advance profited from the glacier-friendly climate conditions (cold/wet) between 1814 and 1822, which led to the first glacier advance in the nineteenth century, at around 1820 (Zumbühl, 1980; Pfister, 1984). The Alpine glaciers were already in an advanced position at the beginning of the last LIA advance. During the Maunder Minimum at AD 1645–1715 (Eddy, 1976), however, winter temperatures were also very low and the summers rather wet but, because of very dry winters (continental winters) no significant glacier advance occurred (Pfister, 1984).

More generally speaking, on a longer timescale, the historical data (Wanner *et al.*, 2000: Figure 3) indicate that, in comparison with the last century, the LIA between 1300 and 1850 coincided with wetter summers and colder winters, i.e., two factors favourable to glacier extension and higher lake-levels.

The weather conditions in Europe may be affected, at least partly, by mecanisms associated with the North Atlantic Oscilllation (NAO) (Hurrell, 1995; Schmutz and Wanner,

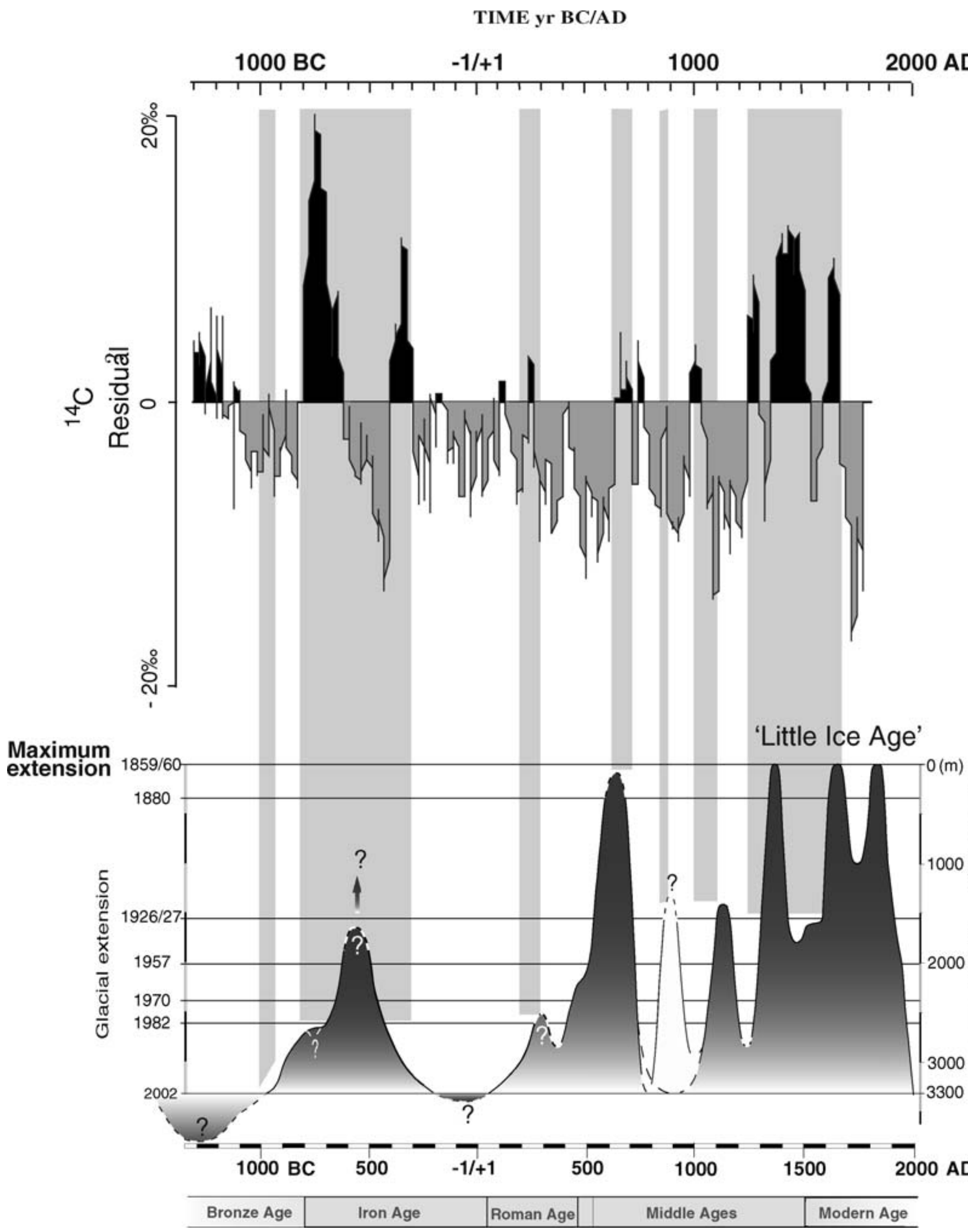

Figure 6 Comparison between the fluctuations of the Great Aletsch glacier (this study) and variations in atmospheric residual ^{14}C (after Stuiver *et al.*, 1998)

1998). The NAO causes periods of predominantly zonal (warm/wet, positive NAO index) or meridional (cold/dry, negative NAO index) flow regimes in northern and southern Europe. It influences the Alps in a complex manner because they are situated in a transition zone between north and south. Thus, the beginning of the impressive loss of the alpine glaciers since the last glacial maximum *c.* 1850–60 was a result of zonal circulation (positive NAO index), with an appropriate temperature rise in the alpine region of 0.6–1°C (Wanner *et al.*, 1998) but rather low precipitation amounts over the Alps (Wanner *et al.*, 2000). In contrast, the meridional mode of the NAO (negative index) dominated, together with low solar activity, the continental dry and cold phase of the Maunder Minimum (Wanner *et al.*, 1995; Luterbacher *et al.*, 2000). Recent investigations have also suggested a possible relation between NAO and varying solar activity (Boberg and Lundstedt, 2002; Gimeno *et al.*, 2003).

Conclusions

(1) In addition to archaeological and historical sources, a set of tree-ring and radiocarbon data allows to recognize that, during the last 3500 years, the Great Aletsch and the Gorner glaciers in the Alps of Valais and the Lower Grindelwald glacier in the Bernese Alps (Switzerland) experienced nearly synchronous advances at *c.* 1000–600 BC, AD 500–600, 800–900, 1100–1200 and, within the LIA, at 1300–1860, while the period 400 BC–AD 400 appears to correspond to a major glacier recession.
(2) Over the past 3500 years, the glacier maximums in the Swiss Alps generally coincided with phases of higher lake level in eastern France (Jura mountains and Pre-Alps) and on the Swiss Plateau.
(3) In agreement with previous studies, a comparison between the fluctuations of the Great Aletsch glacier and the

variations in the atmospheric residual ^{14}C records supports the hypothesis that variations in solar activity were a major forcing factor of climatic oscillations in west-central Europe during the late Holocene.

(4) Detailed historical data indicate that glacier advances since AD 1550 have corresponded to characteristic but varied weather conditions and regimes. On the other hand, the general trend shown by the historical data suggests that, in comparison with the last century, the LIA coincided with cooler winters and wetter summers favouring both glacier advances and higher lake-levels.

Acknowledgements

This study profited from the authors' stimulating discussions with H. Wanner (University of Bern). Financial support was provided by the French CNRS (Program ECLIPSE) and the European Union (Project ACCROTELM, leader: F. Chambers). The authors express their sincere thanks to J.C. Rougeot for his help with the figure drawings, to J. Olsen for his help with the English language, to K. Nicolussi for his help with dendro dating some tree-ring width sequences, and to the reviewers M. Maisch and W. Tinner.

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