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High-resolution record of environmental changes and tephrochronological markers of the Last Glacial–Holocene transition at Lake Lautrey (Jura, France)

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ABSTRACT: This paper presents the results of a multiproxy investigation including volume magnetic susceptibility (κ), mineral and pollen analyses of Late Glacial sediments from Lake Lautrey (Jura, France). Small-scale lithological variations have been identified with high stratigraphic resolution in order to establish lithostratigraphic correlations between cores. κ measurements, combined with mineralogical analyses, provide information on past sedimentary processes. This combined approach reflects major changes in terrestrial habitats and soil processes which may relate to the climatic events characterising the Late Glacial climatic warming and cooling phases. During warm intervals, the record indicates increased lake productivity via carbonate precipitation and decreased input of detrital material. In contrast, cooler intervals show reduced lake productivity, catchment area instability and increased detrital inputs. Several short interruptions in reforestation and in soil stabilisation can be identified and linked with abrupt colder events occurring through the Bølling. A general trend of warming is recorded from the coldest part of the Younger Dryas. Three tephra layers were also detected. The mineral composition analyses show that the upper tephra layer corresponds to the Laacher See eruption (Eifel, Germany) while the lower ones may relate to the volcanic activity of the Chaîne des Puys (Massif Central, France) around 13 000 cal. yr BP. These two events, recognised for the first time outside the Massif Central region, may provide additional chronostratigraphic markers for the Late Glacial sedimentary records of the Jura mountains and northern Alps. Copyright © 2004 John Wiley & Sons, Ltd.

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KEYWORDS: magnetic susceptibility; Late Glacial; lake sedimentation; palaeoecological changes; tephrochronology.

Introduction

Studies of ice-sheet oxygen isotopic records have made it possible to detect the successive abrupt climatic changes of the Last Glacial–Interglacial Transition in the north Atlantic region (Björck *et al.*, 1998; Johnsen *et al.*, 2001). However, the resulting ecological modifications have hitherto only been recognised at a regional scale. It is therefore difficult to establish the chronological link between these two facets of environmental evolution (Ammann and Oldfield, 2000). This is why tephra deposits appear to be of major interest in establishing time markers for the Last Glacial–Interglacial Transition (Lowe, 2001).

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Various proxies such as pollen, chironomid assemblages or oxygen isotope ratios have been used for quantitative reconstruction of past temperatures, and also to analyse biological variations due to climatic changes (von Grafenstein *et al.*, 1999; Ammann *et al.*, 2000; Millet *et al.*, 2003). Sedimentary parameters also constitute key proxies for describing and analysing geo-ecosystem responses to these climatic shifts (Brauer *et al.*, 1999a; Stockhausen and Zolitschka, 1999). Measurements of magnetic susceptibility are a useful tool for detecting environmental changes on a large scale (Thouveny *et al.*, 1994; Kukla *et al.*, 2002). Catchment area erosion, soil processes, volcanic activity and lake productivity can result in different magnetic mineral concentrations in lake sedimentary sequences (Sandgren and Snowball, 2001). Recent work based on several European carbonaceous lakes show that magnetic susceptibility measurements reflect detrital inputs in response to both climatic cooling and vegetation cover changes during the Last Glacial–Interglacial transition (Wessels, 1998; Nolan *et al.*, 1999). Such analyses can also rapidly detect micro tephra deposits in sedimentary sequences (van den Bogaard *et al.*, 1994).

This paper presents the results of surface scanning magnetic susceptibility, mineralogy and pollen analyses from several Late Glacial (14 700–11 500 cal. yr BP) sediment cores from Lake Le Lautrey (Jura, France). Volume magnetic susceptibility (κ) was used to correlate sediment sections from different parts of the lake basin and to detect tephra layers. The aims were to provide detailed information on the spatial distribution of the sediments in the basin and to analyse the time variability of the sedimentary accumulation. Variations in sedimentary processes have been evaluated in terms of environmental changes in relation to mineralogical modifications and local vegetation history. These results were subsequently compared to those of lake-level variations and chironomid fauna analyses (Magny *et al.*, 2002; Millet *et al.*, 2003).

Study area

Lake Lautrey (46°35′14″ N, 5°51′50″ E; 788 m a.s.l) is located in the French Jura Mountains at the limit between the calcareous plateau and the folded High Jura Chain (Fig. 1(a)). The lake basin (500 × 200 m) is situated in a small catchment area (2 km^2) underpain by Jurassic and Cretaceous limestone. At present, it consists of a peaty depression mostly filled up with Quaternary lacustrine deposits and a residual pond. The geophysical survey of the substrate morphology has revealed three lake sub-basins (Bossuet *et al.*, 2000; Fig. 1(b)). The pre-lacustrine topography shows a gentle slope in the northwestern part of the basin and a steep slope in the southeastern part formed by the Cretaceous cliff. In the deepest zones, the maximal water depth during the Late Glacial period reached 12 m (Fig. 1(b)).

Methods and materials

Field methods and materials

Six sedimentary sequences were obtained with a strengthened Russian corer (providing split core 10 cm in diameter and 1 m in length), (L1 to L6; Fig. 1(b) and 1(c)). Lengths of the cores were, respectively for L1 to L6, 170 cm, 370 cm, 640 cm, 290 cm, 840 cm and 460 cm. The stratigraphy shows a typical Jurassian infilling sequence with minerogenic sediments at the base (laminated silty-clay and two clay layers interrupted by silty-clay lacustrine marl), lacustrine marl in the middle and organic sediments at the top (peat formations; Fig. 1(c)). Previous studies on Lake Lautrey—manual core drilling exploration (44 cores) and pollen stratigraphic analyses—have shown that silty-clay and calcareous clay deposits correspond to the Last Glacial–Interglacial Transition while the upper lacustrine marl and peat deposit represent Holocene sedimentation (Bossuet *et al.*, 2000).

Laboratory methods

The cores were described in the field and the lowermost levels of the six sequences, corresponding to the Last Glacial–Interglacial Transition, were selected for further analysis. These cores were placed in half PVC tubes, wrapped in plastic film and stored at 4 °C in a cold room. Prior to κ measurements and sub-sampling, split half core surfaces were carefully cleaned and described. Mineralogical and pollen analysis were undertaken on core L6 which had the highest temporal resolution.

κ was performed to aid the visual correlation between the different cores and to detect the finest lithological variations induced by in-wash of non-carbonaceous minerogenic material. κ was measured with a MS2E1 surface scanning sensor from Bartington Instruments. This sensor is well adapted to measure the volume magnetic susceptibility of split cores with fine resolution (Lees *et al.*, 1998; Nowaczyk, 2001; Sandgren

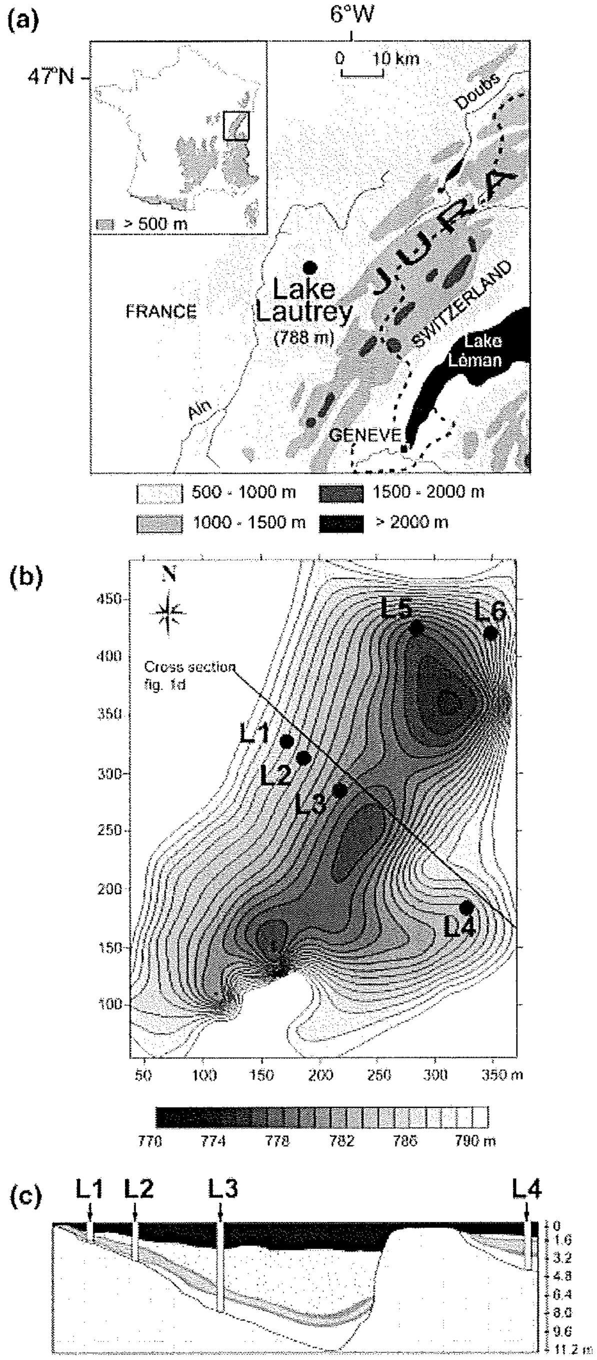

Figure 1 (a) Location of Lake Lautrey (Jura, France). (b) Pre-lacustrine topography of the basin and location of boreholes. (c) Sedimentary cross-section of the lake infilling

Table 1 AMS radiocarbon dates of core L6 of Lake Lautrey (Jura, France). Radiocarbon dates calibrated with CALIB 4.3 (Stuiver *et al.*, 1998)

Depth (cm)	Radiocarbon age BP	Calibrated age 1σ cal. yr BP	Calibrated age 2σ cal. yr BP	Laboratory reference	Material
295–296	10 000 ± 40	11 553 (11 400) 11 261	11 917 (11 400) 11 256	VERA 1716	Twigs
304–305	10 140 ± 35	11 946 (11 720) 11 604	12 281 (11 720) 11 441	VERA 1715	Twigs
323–324	10 960 ± 45	13 117 (12 984) 12 890	13 153 (12 984) 12 656	VERA 1724	Twigs
325–326	11 100 ± 45	13 159 (13 132) 13 000	13 190 (13 132) 12 682	VERA 1725	Twigs
365–366	11 575 ± 45	13 800 (13 482) 13 440	13 835 (13 482) 13 316	VERA 1726	Twigs
383–384	11 695 ± 45	13 828 (13 676) 13 488	15 091 (13 676) 13 441	VERA 1727	Twigs
397–398	11 825 ± 45	14 013 (13 827) 13 643	15 204 (13 827) 13 586	VERA 1728	Twigs
416–417	12 170 ± 60	15 177 (14 123) 14 090	15 400 (14 123) 13 836	POZ 4496	Twigs
425–426	12 430 ± 45	15 375 (14 350) 14 247	15 516 (14 350) 14 138	VERA 1729	Twigs
444–445	12 590 ± 45	15 476 (14 619) 14 354	15 591 (14 619) 14 222	VERA 1730	Twigs

and Snowball, 2001). Measurements were made at 5-mm intervals.

X-ray diffraction analyses were carried out to obtain information about the nature of the non-carbonaceous minerogenic material (X'Pert Philips Diffractometer with a cobalt anticathode). Sediments with the highest κ values were further analysed by photonic microscope observations, scanning electron microscope (Jeol 5600 with X EDS Fondis-99 microanalysis) and X-ray diffraction analyses (with a SCINTAG XRD 2000 Diffractometer) in order to obtain more detail and precision on mineralogical characterisation. Chemical analyses were conducted on glass shards and phenocrysts using an electron microprobe (EDS-WDS) from the 'Institut de Physique du Globe' (University of Paris 7).

Pollen analyses were carried out at 2-cm intervals in order to study upland vegetation changes. The sediment samples were treated by standard methods, including HCl, HF, NaOH, and acetolysis. The number of pollen grains counted per sample was up to 500. Pollen percentages were based on the sum of arboreal pollen (AP) and non-arboreal pollen (NAP), excluding spores. Simplified pollen diagrams with main taxa only (*Juniperus, Betula, Pinus, Poaceae, Artemisia* and other herbs) are presented here.

Chronology

The chronological framework of the sedimentary sequence is based on:

- 10 AMS radiocarbon dates of terrestrial plant macrofossils from core L6 (15 500 to 11 500 cal. yr BP; Table 1). Radiocarbon dating was carried out on 2–3 cm^3 of fresh material sieved at 125 μm. In order to avoid hard-water effect, terrestrial macrofossils (twigs) were selected and sent to Vienna (VERA) and Poznan (POZ) laboratories. The radiocarbon dates were calibrated to produce a calendar-year chronology using CALIB 4.3 (Stuiver *et al.*, 1998). An age/depth model was established for core L6 by linear interpolation of the calibration data points corresponding to 10 radiocarbon ages and the LST (Fig. 2).
- tephrochronology using the millimetric ash layer of Laacher See Tephra (LST) previously identified in Lake Le Lautrey (Bossuet *et al.*, 1997) and also recognised in cores L2 (depth = 297.5 cm), L3 (588.5 cm), L5 (793 cm) and L6 (344.5 cm). It has been dated to 12 880 cal. yr BP from varve counting in the Meerfelder Maar (Brauer *et al.*, 1999b; Zolitschka *et al.*, 2000).

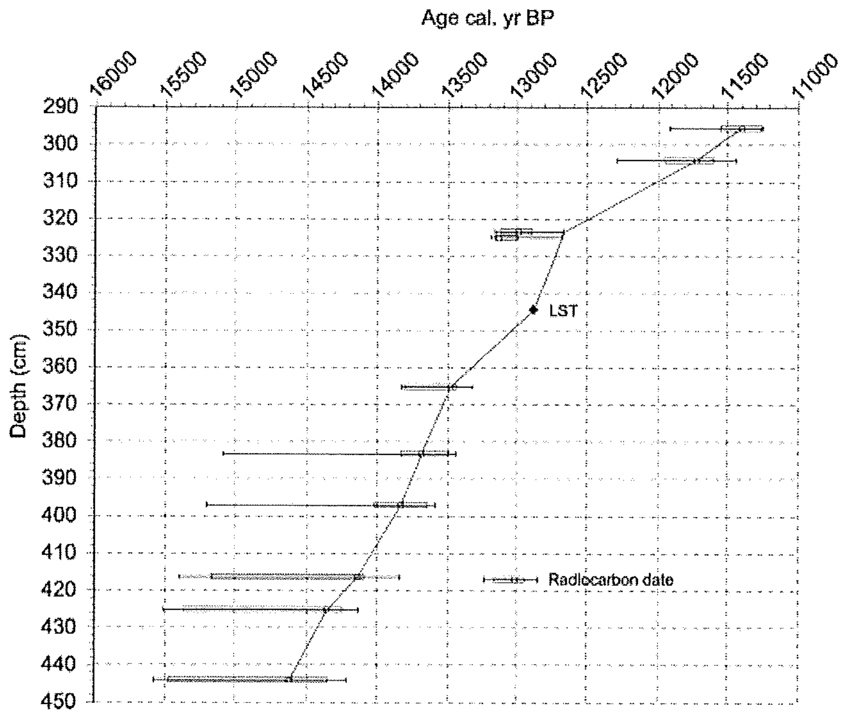

Figure 2 Time/depth model of core L6 based on ^{14}C age ranges (Table 1)

— local assemblage pollen zones (LPAZ) from core L6 based on detailed Last Glacial–Interglacial Transition pollen stratigraphy which has been established for the neighbouring Swiss Plateau using over 90 AMS radiocarbon dates (Ammann and Lotter, 1989). A similar pollen stratigraphy has been recognised in the Jura Mountains and Northern French Pre-Alps over the period 15 000–11 000 cal. yr BP (Wegmüller, 1966; Gaillard, 1984; de Beaulieu *et al.*, 1994; Wohlfarth *et al.*, 1994; Bégeot *et al.*, 2000).

Results

Lithology and magnetic susceptibility

Sedimentary changes and the major variations in κ values allow us to define five common stratigraphic units among the different cores (Figs 3 and 4; Magnetic susceptibility zones: MSz-1 to MSz-5). MSz-1 is characterised by dark-grey clay and high κ values (up to 50 10^{-6} SI units). The abrupt decrease of κ values delimits the onset of zone MSz-2 which also corresponds to a succession of yellow-grey calcareous clay and grey-green silty clay lacustrine marl deposit. Through this zone there is a progressive decrease of κ values but sporadic light increases, connected with yellow-grey calcareous clay sedimentation, are also noted. The zone MSz-3 is characterised by low variability, lower values of κ and yellow or pink lacustrine marl deposits. MSz-4 is marked by an abrupt increase of κ values and grey clay sedimentation. A general decrease of κ values is recorded within this zone which can be connected with a progressive enhancement of calcareous components from grey clay to silty clay lacustrine marl. Lower κ values are recorded again in zone MSz-5 and correspond to beige lacustrine marl accumulation. Three distinct peaks are observed in zone MSz-3 (LST, LAUT1 and LAUT2; Fig. 4). The highest one corresponds to the ash layer of the Laacher See eruption, clearly recorded in the lithology (LST). The LAUT1 and LAUT2 peaks do not correspond to visually detectable lithological changes.

κ variations appear closely connected with the major lithological changes in all cores. The higher κ values of MSz-1 and MSz-4 are related directly to an increase of detrital input (clay, quartz and plagioclase; Figs 3, 5 and 6). On the other hand, κ values are lower when sedimentation is dominated by autochthonous calcareous deposits (lacustrine marl of MSz-3 and MSz-5). Thus, MSz-2 may be considered as an intermediate zone; the progressive decrease of MS values reflects the transition between allochthonous and autochthonous sediments. κ measurements indicate the overall concentration of magnetic minerals in the sediments; since Jurassic limestone lacks ferromagnetic minerals, most of these must be derived from topsoil layers or atmospheric sources such as dust transported by storms or volcanic ash. Aeolian silt from the Alps forms non-calcareous soil on the Jura Mountains which can be an important source of allochthonous material such as quartz, plagioclase and magnetic minerals. Thus, κ values reflect sediment mobility over the catchment area and erosion processes (Thompson and Oldfield, 1986; Stockhausen and Zolitschka, 1999).

The stratigraphic correlations between the different cores highlight significant variations in sedimentation rates within the basin which can be connected with its morphology. Sedimentation is relatively less important towards shore (core L1; Fig. 3) and in the deepest part of the basin (cores L3 and L5). The maximum sedimentation rate occurs in core L6 which is located on a steep slope at the northeast of the basin. Clay and calcareous clay deposits of zone MSz-4 reach the maximum thickness in the deepest part of the basin (cores L3 and L5), whereas lacustrine marl deposits of zones MSz-2 and MSz-3 are better developed in cores L2, L4 and L6, located in the middle of the basin. This is in agreement with the carbonate precipitation belt zone which characterises Jura lakes (Magny, 1998). That the sediment record from core L1, near the shore, is the less developed may be a result of lake-level variations. Nevertheless, all cores have the same sedimentary succession and the outline of the κ curves appears similar (Fig. 4). This suggests the absence of hiatuses despite high variability within each zone among the different cores' κ profiles. These differences are particularly marked in LST, LAUT1 and LAUT2, detectable only in the deepest part of the basin, above 400 cm depth (L3, L5 and L6) where sedimentation is continuous and not disturbed.

Palynology

In the simplified pollen diagram of core L6 (Fig. 5), significant frequency variations of dominant taxa (*Pinus, Betula, Juniperus, Artemisia* and the AP/NAP ratio) enable us to identify six LPAZ; these can be related to the regional biozones of Oldest Dryas, Bølling, Older Dryas, Allerød, Younger Dryas and Preboreal, observed in the Jura Mountains and on the Swiss Plateau (Gaillard, 1984; Ammann and Lotter, 1989; Ammann *et al.*, 1994; Wohlfarth *et al.*, 1994). The lowermost LPAZ 1 is dominated by high percentages of herbaceous pollen, mainly *Artemisia*, Poaceae and other heliophilous taxa. Pollen concentrations are very low. This assemblage indicates an open herbaceous landscape and can be correlated with the Oldest Dryas biozones (de Beaulieu *et al.*, 1994). Low *Pinus* values (less than 25%) occur as a result of long-distance transport. Around 12 590 $\pm$ 45 yr BP, the increase of *Betula* and *Juniperus* in LPAZ 2 characterises the first part of the Bølling period. The increase of *Juniperus* (22.6%) corresponds to the extension of areas covered with a Juniper scrub. The second part of the period corresponds to the establishment of *Betula* woodland as indicated by high pollen percentages and concentrations (LPAZ 3a). Three successive *Betula* peaks are recognised along with a decrease in herbaceous pollen (LPAZ 3a and 3b). This may reflect successive phases of expansion and regression in vegetation development. At 420 cm, the decline in *Betula* pollen percentages, while those of *Juniperus*, Poaceae *and Artemisia* increase, marks the Older Dryas biozone (LPAZ 3b; de Beaulieu *et al.*, 1994; Bégeot *et al.*, 2000). Subsequently the increase of *Pinus* pollen percentages indicates that pines penetrated rapidly into the birch woodland and became the dominant tree species during the Allerød (LPAZ 4). There is a decline in percentages of Poaceae and heliophilous herbs while arboreal pollen dominates. The last part of the Allerød is characterized by a relative decrease in *Pinus* percentages together with an increase in *Betula*, Poaceae, *Artemisia* and some other NAP taxa. Around 10 960 $\pm$ 45 BP, the rise of NAP pollen percentages, mainly *Artemisia and* Poaceae, indicates the reduction of the pine–birch forest and the expansion of open areas, also recorded in the pollen concentrations (LPAZ 5). This zone corresponds to the beginning of the Younger Dryas. *Juniperus* pollen is also common during the Younger Dryas but less important than at the beginning of the Bølling. At 298 cm, the increase in *Betula* pollen, the increase of pollen concentrations and the decrease in Poaceae, *Artemisia* and most of the other heliophilous taxa, indicate the spread of birch–pine woodland and the Younger Dryas/Preboreal transition (LPAZ 6; Fig. 5).

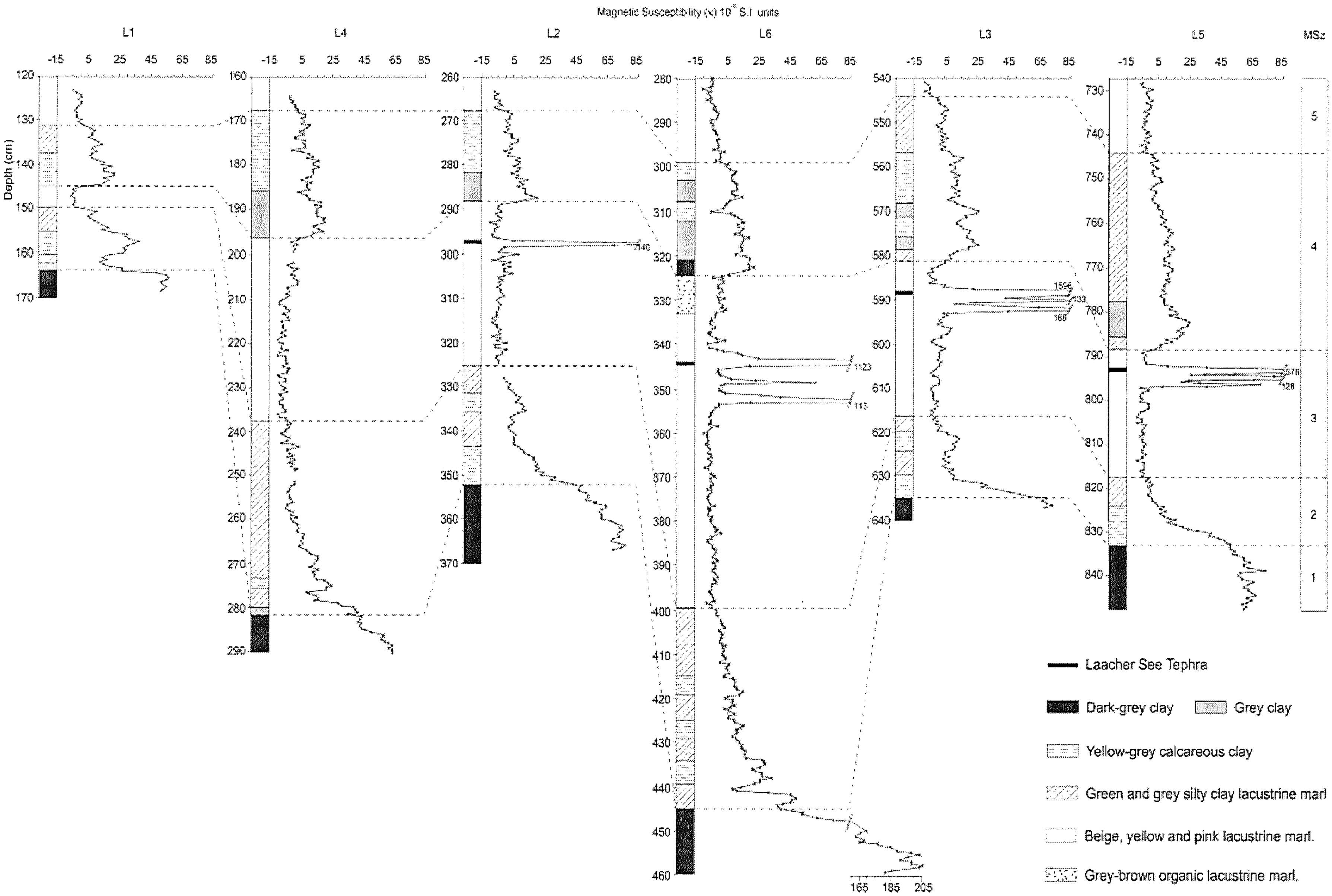

Figure 3 Lithostratigraphy and volume magnetic of the sediment records from cores L1, L2, L3, L4, L5 and L6 of Lake Lautrey (Jura, France). Cores log are classified relating to their position in the basin from the shore to the deepest part. Determination of the magnetic susceptibility zones (MSz)

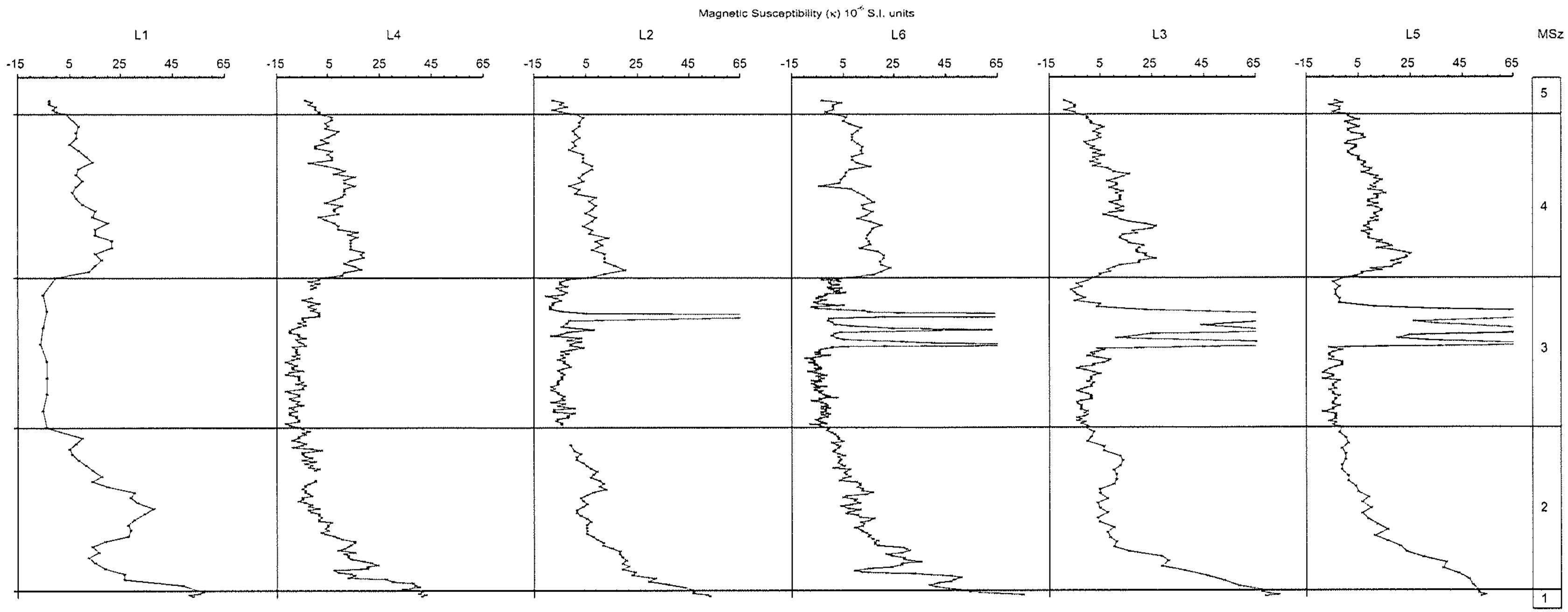

Figure 4 Volume magnetic susceptibility logs from cores L1, L2, L3, L4, L5 and L6 plotted along a common depth scale (cf. Fig. 3)

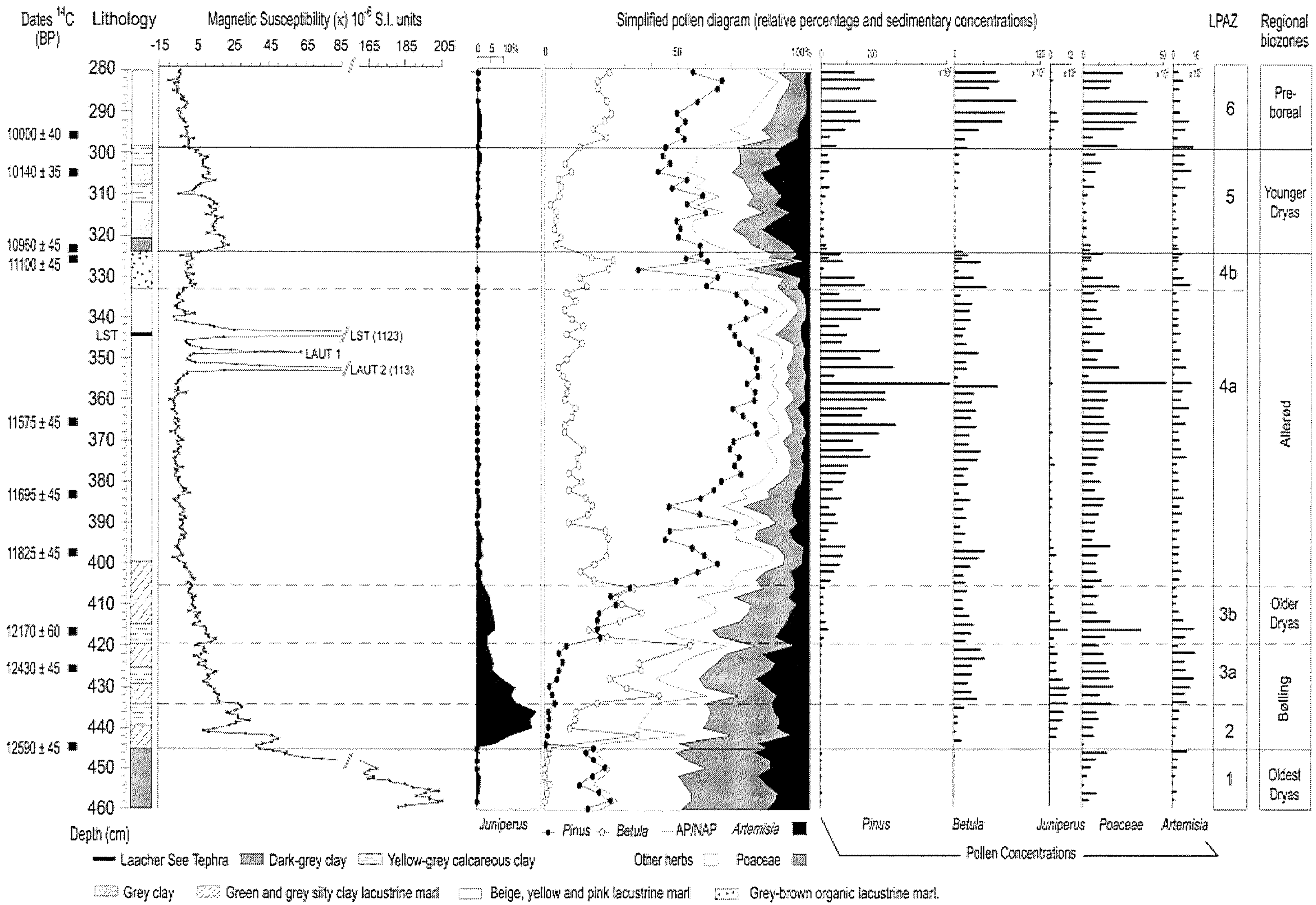

Figure 5 Radiocarbon chronology, lithostratigraphy, volume magnetic susceptibility and a simplified pollen diagram (relative percentage and sedimentary pollen concentrations) from core L6 of Lake Lautrey (Jura, France). Correlations with regional biozones

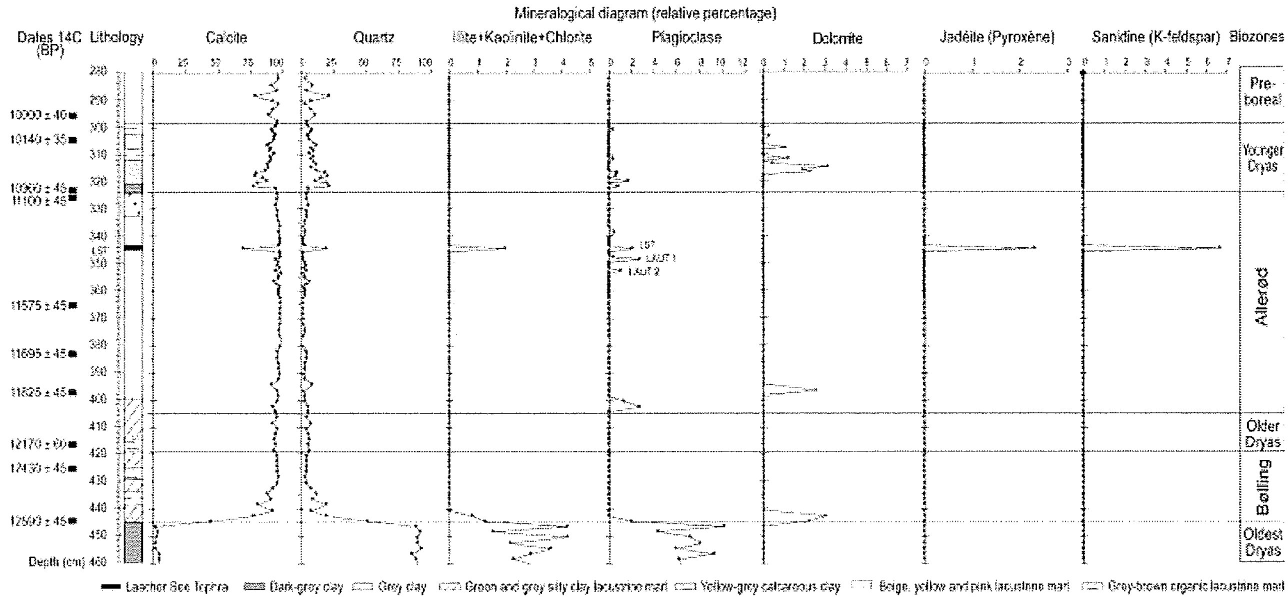

Figure 6 Radiocarbon chronology, lithostratigraphy and mineralogy diagram (relative percentage) from core L6 of Lake Lautrey (Jura, France)

The radiocarbon date of 10 000 ± 45 yr BP confirms this interpretation.

Mineralogy

The mineralogy of the L6 core sediments is mainly dominated by calcite, sometimes associated with quartz (Fig. 6). Semi-quantitative results indicate that clayey minerals (mainly illite, kaolinite, chlorite) and feldspars (mainly plagioclase) are present in very low quantities. The Oldest Dryas is characterised by the predominance of quartz, the presence of feldspars and clay minerals, and low calcite content. Quartz appears as silty grains with wind abrasion marks in microscopic observations. The Oldest Dryas–Bølling transition corresponds to a strong decrease in the amount of quartz, feldspars and clay minerals, and to an increase in calcite. During the Bølling–Allerød period, four peaks in plagioclase (in LAUT2, LAUT1, LST samples and around 402 cm deep) and one increase in quartz, jadeite and sanidine (LST sample) are observed. The Allerød–Younger Dryas transition corresponds to an increase in quartz and feldspars together with a relative decrease of calcite. The amount of quartz slightly decreases towards the end of the Younger Dryas.

The mineral content of samples with highest κ values (LST, LAUT1 and LAUT2; Fig. 5) consists of volcanic glass shards and magmatic minerals, together with calcite, quartz and clay minerals (Table 2). Volcanic glass shards and magmatic minerals were not observed in the other sediment samples from core L6. The three peaks of κ detected in cores L2, L3, L5 and L6 attest to the occurrence of three tephra layers in the Lake Lautrey sequence. The most recent tephra layer, LST, contains unaltered brown hornblende, jadeite (pyroxene), hauyne, sanidine (K-feldspar), plagioclase and sphene, which are magmatic minerals characteristic of the Laacher See Tephra composition (Table 2; Fig. 7; van den Bogaard and Schmincke, 1985). The second tephra layer, LAUT1, contains magmatic glass shards, unaltered Ca–Na plagioclase (albite–anorthite), augite and small amounts of olivine. The earliest tephra layer, LAUT2, consists of magmatic glass shards, unaltered Ca–Na plagioclase (albite–anorthite), sphene, spinel, olivine and micas (Table 2; Fig. 7). In all three layers, diagenetic pyrite, clayey mineral and zeolites have also been observed (Tables 2 and 3). Basaltic volcanic glasses are very sensitive to weathering that occurs after the deposit of the tephra. The consequences of glass weathering are hydration of glass and neoformation of zeolites and clay minerals (especially bentonite, a mixture of clay minerals, mostly montmorillonite; e.g. Kamei *et al.*, 2000; Dahlgren *et al.*, 1999; Lowe, 1986). Microprobe chemical analyses of the less weathered glass shards from LAUT1 and LAUT2 show high Al_2O_3 content and high variability of K_2O content which prove that these glass shards are weathered. Nevertheless, on a $SiO_2/Na_2O + K_2O$ diagram, the mineralogical analyses of LAUT1 and LAUT2 magmatic phenocrysts, which resist weathering since they are crystalline structures, show the trachy-andesitic composition of these tephra deposits (Table 3).

Table 2 Mineralogical composition of air-fall tephra layers LST, LAUT1 and LAUT2 from petrographic and SEM observations, and X-ray diffraction analyses

Tephra	Petrographic observations	MEB	X-ray diffraction
LST	Volcanic glass, Sphene, Pyroxene, Hauyne, Feldspars, Brown hornblende	Volcanic glass, Feldspars, Pyrite, Pyroxene, Weathered volcanic glass, Clay minerals	Jadeite, Sanidine, Plagioclase, Clay minerals
LAUT1	Volcanic glass, Feldspars, Olivine	Volcanic glass, Feldspars, Pyrite, Weathered volcanic glass, Clay minerals	Anorthose, Clay minerals
LAUT2	Volcanic glass, Feldspars, Micas	Volcanic glass, Feldspars, Pyrite, Weathered volcanic glass, Clay minerals	Albite, Micas, Clay minerals

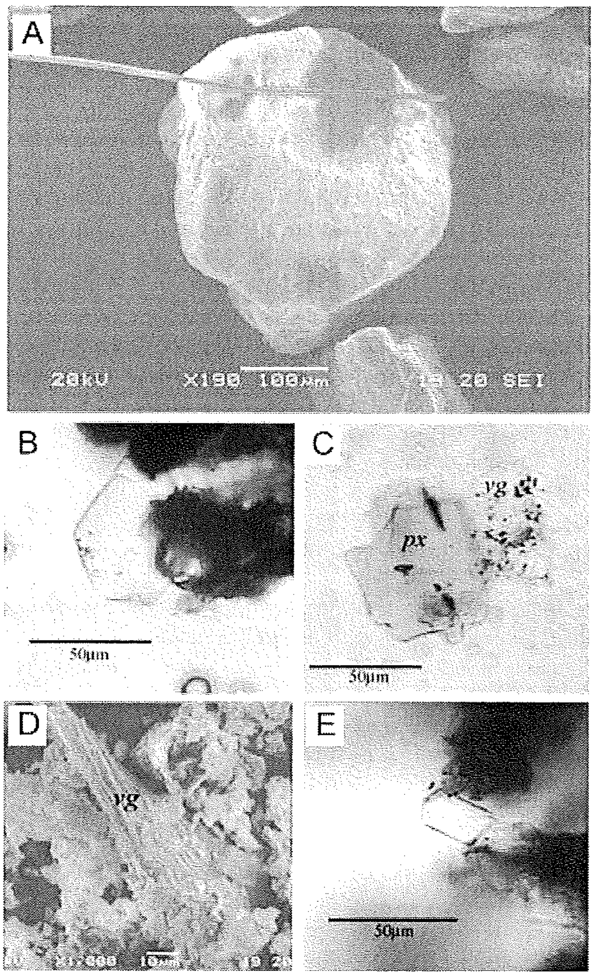

Figure 7 Photographs of minerals from the tephra layers of Lake Lautrey (Jura, France). A, Quartz with aeolian marks (SEM); B, mica sheet from LAUT2 tephra; C, pyroxenes (px) and volcanic glass (vg) particles from LST; D, volcanic glass shard (vg) from LST (SEM); E, plagioclase from LAUT2 tephra

Discussion

Late Glacial environmental changes

According to κ, pollen and mineralogical analyses from core L6, several Late Glacial phases are identified at Le Lautrey (Figs 5 and 6).

During the Oldest Dryas (LPAZ 1), environmental parameters indicate an open landscape favouring detrital input, particularly by aeolian transport as testified by the presence of wind faceted quartz grains (Fig. 7). The low proportion of biogenically precipitated carbonate reflects poor biological activity as a result of low temperature and low inputs of allochthonous organic matter in relation to a treeless environment and a cool climate.

Then, the increase in calcite attests to higher biological activity (Fig. 6) favouring the development of a large marginal calcareous platform which induces the high sedimentation accumulation rate observed in cores L2, L4 and L6. The first stage of *Juniperus* and *Betula* colonisation corresponds to the beginning of the Bølling–Allerød Interstadial (LPAZ 2–3–4), characterised by an abrupt climatic warming (Fig. 5; Gaillard, 1984; Ammann *et al.*, 1994; Schwander *et al.*, 2000; Renssen and Isarin, 2001) and a lake-level lowering on a regional scale (Magny, 2001). The low *Betula* pollen concentration indicates an open birch forest. The successive increases of pollen concentration reflect progressive colonisation interrupted by regressive short periods at 440 and 429 cm (LPAZ 2–3a). The progressive decrease trend in κ values is also sporadically interrupted by short increases, corresponding to clayey sedimentation and to the *Betula* pollen percentages decreases. The relationship observed between the *Betula* pollen percentages curve and κ variations show the conspicuous role of this species in the spread of the forest and in soil stabilisation (Wick, 2000). Such interruptions in vegetation dynamics may reflect high climate variability with rapid cooling events occurring throughout the Bølling period. They may be linked to the Intra-Bølling Cold Phases (IBCPs) recognised in previous pollen records or lake-level reconstructions from the Jura Mountains (Bégeot *et al.*, 2000; Magny and Richoz, 2000). At Gerzensee (Swiss Plateau) two cold episodes during this period have also been identified by oxygen-isotope and pollen studies (Lotter *et al.*, 1992). At 420 cm (LPAZ 3b), the non-arboreal pollen (NAP) concentration increases and the re-expansion of juniper is concomitant with a relative increase of κ values and clayey deposits. These records reflect the cooling period corresponding to the Older Dryas regional biozone or to the Aegelsee oscillation (Lotter et al., 1992). This is confirmed by a radiocarbon date (12 170 ± 60 yr BP) obtained at 416–417 cm (Fig. 5; Björck, 1984; Lowe *et al.*, 1994; Lotter *et al.*, 1992).

The rapid increase in *Pinus* pollen concentration and percentage marks a second step in vegetation dynamics characterising the Allerød period (LPAZ 4a). Pioneer taxa, such as *Betula* and *Juniperus*, give way to a *Pinus*-dominated forest with a slight decrease in open habitats, better adapted to the warmer climate. NAP and κ values reach their minimum. The sediments

Table 3 Results of microprobe chemical analyses of air-fall tephra layers LAUT1 and LAUT2

	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	Total
LAUT1											
Glass shards	51.61	0.23	20.70	1.70	0.06	0.48	6.78	5.25	1.23	0.23	88.26
Albite	69.34	0.00	19.09	0.03	0.08	0.01	0.00	11.53	0.07	0.03	100.19
Plagioclase	56.48	0.24	26.04	0.83	0.02	0.12	9.51	5.50	0.59	0.08	99.41
K-feldspars	62.29	0.21	17.16	0.11	0.00	0.01	0.19	0.92	14.46	0.01	95.35
Augite	48.97	0.43	5.67	13.36	0.42	13.10	11.96	0.66	0.18	0.05	94.79
LAUT2											
Glass shards	58.59	0.00	17.66	0.02	0.00	0.01	0.96	9.73	0.09	0.03	87.09
Albite	67.17	0.00	19.37	0.05	0.00	0.01	0.46	10.76	0.09	0.00	97.91
Plagioclase	53.86	0.05	28.32	0.56	0.00	0.09	11.15	4.88	0.39	0.01	99.32
K-feldspars	64.01	0.00	18.71	0.00	0.00	0.00	0.07	0.81	15.26	0.04	98.90
Sphene	23.86	33.27	1.10	1.23	0.00	0.01	25.31	0.00	0.00	0.13	84.91
Spinel	0.39	24.48	14.58	20.99	0.55	7.32	0.63	0.03	0.04	0.13	69.13
Olivine	41.50	0.00	0.02	4.70	0.27	52.67	0.04	0.02	0.00	0.02	99.24

are dominated by autochthonous carbonates. At the end of the Allerød biozone (333 cm), the slight increase in κ values correlates with an increase of *Betula* and herbs (LPAZ 4b). The sedimento become more organic. At first sight, the NAP increase seems to reflect a landscape opening, which may be related to the climatic cooling at the onset of the Younger Dryas. However, this environmental change is recorded by the end of the Allerød (before 13 190–12 682 cal. yr BP; Table 1). It corresponds to a lake phase of low lake-water level, well recorded by peaks of organic matter and oncolithes (Magny *et al.*, 2002) and to the extension of a marginal shallow-water zone revealed by chironomid faunal assemblages (Millet *et al.*, 2003). These drier soil conditions around the lake would have favoured *Betula* colonisation in the vicinity.

The Younger Dryas biozone (LPAZ 5; Figs 5 and 6) is clearly delimited by an important decrease of tree pollen concentration, NAP percentages up to 40% and a sharp increase in κ followed by a continuous decrease. This environmental change, occurring around 13 190–12 682 cal. yr BP, is visually identified by the lithological change, from biochemical (beige lacustrine marl) to detrital sediments (dark-grey clay, with an increase in quartz). Pollen assemblages and the absence of tree macrofossils suggest that the forest limit was situated at a lower altitude (Bégeot, 2000). A simultaneous event is recorded in the chironomid assemblages which indicate a drop in temperature within this zone (Millet *et al.*, 2003). During the Younger Dryas, the increase in detrital inputs (quartz and plagioclase; Fig. 6) may result from both erosion of soils rich in aeolian minerals, inherited from the Last Glacial Maximum, or from an increase in atmospherically transported sediments, as revealed by higher dust concentrations in the Greenland atmosphere (Mayewski *et al.*, 1993). The Younger Dryas has also been described as having higher frequencies of stormy events and an expansion of dry open areas on a global scale, leading to reduced vegetation cover (Bond *et al.*, 1997). In the neighbouring peat bog of la Gruère (Swiss Jura Mountains), Shotyk *et al.* (2001) have recognised important deposits of soil dust around 10 590 ^{14}C yr BP. Grain size and mineralogy indicate an increase in wind strength and a change in source area. One supplementary argument, important in connecting this result with the κ signal at Lake Lautrey, is that the lithogenic element Sc, an indicator of soil dust increase, may originate particularly from ferromagnetic minerals (Shotyk *et al.*, 2001). Most of the lacustrine sequences from the Jura and Alpine Mountains or the Swiss Plateau show the same common increase of detrital minerogenic inputs during this period as a result of accelerated soil erosion (Wohlfarth *et al.*, 1994). In contrast with the Oldest Dryas biozone, the persistence of the high production of biogenic carbonate lake marl suggests that summer temperatures were sufficient for carbonate precipitation and biologic activity. AGCM experiments indicate that the Younger Dryas climate was particularly marked by a change in precipitation which was concentrated during the summer season (Renssen and Isarin, 2001) which favours lake-levels increases at a regional scale (Magny, 2001)

Lithology shows a subdivision of the Younger Dryas (Fig. 3) into (i) an early phase with grey clay facies, characterised by predominant allochthonous deposits, corresponding to the highest κ values, and (ii) a later phase with calcareous clay facies, marked by an increase in biological activity favouring carbonate precipitation. A similar division has been described in lake and fluvial sequences in Germany, where the Younger Dryas can be subdivided into an early phase, characterised by periglacial processes in elevated areas, while rainwater-driven discharge developed later on (Brauer *et al.*, 1999b; Andres *et al.*, 2001). This subdivision is related to climatic conditions: the first phase is colder and wetter while the latter is warmer and drier (Lotter *et al.*, 1992); furthermore, these characteristics are also recorded by other vegetation and lake-level reconstructions in Western Europe (Isarin, 1997; Bos, 2001; Magny *et al.*, 2003). At Le Lautrey, in the upper part of the Younger Dryas, the slight increase of *Betula* and concentrations of other lake may reveal reforestation starting before the end of this period. This can favour soil stability (Fig. 5). All the κ curves from the six cores (L1 to L6) show a general decreasing trend towards the end of this period (Figs 3 and 4). This pattern reflects the progressive decrease in detrital input in lake sedimentation, also revealed by mineralogical assemblages and decrease in quartz and plagioclase content. The Younger Dryas (GS-1) is characterised in the Swiss Plateau oxygen isotope records (von Grafenstein *et al.*, 1999; Schwander *et al.*, 2000) and at Lake Le Locle (Magny *et al.*, 2001) by a general trend to a slightly warmer climate.

Source and age of the tephra layers

The mineralogical observations and analyses confirm previous results concerning the occurrence of the Laacher See Tephra in the Le Lautrey sequence (Bossuet *et al.*, 1997). The deposits of two older tephra layers have never been recognised before in the lakes of the Jura and northern Alps. Their mineral composition corresponds to a trachytic eruption. Around 12 000–11 000 BP, four distincts volcanic centres or provinces were active in western Europe and produced tephra deposits (Davies *et al.*, 2002): the Laacher See Tephra which was identified; the Icelandic tephra (Vedde Ash and Borrobol Tephra; Davies *et al.*, 2003); the Italian Volcanic Province (Phlegrean Fields, Vesuvio, Etna); and in the Massif Central, Le Puy de la Nugère which is the only known active trachy-andesitic volcano. The mineralogy of the tephra originating from the Puy de la Nugère is dominated by Ca–Na plagioclases (sometimes mantled by K-feldspars; Maury *et al.*, 1980) with clinopyroxenes, spinel and rare olivine (Etlicher *et al.*, 1987; Juvigné, 1987). Our data show that the LAUT1 and LAUT2 tephra are similar to the tephra originating from Le Puy de la Nugère.

Two tephras from Le Puy de la Nugère, slightly older than that of the LST, have been described in the Chaîne des Puys by Juvigné *et al.* (1996) and Vernet and Raynal (2000). The older one (12 010 $\pm$ 150 ^{14}C yr BP, i.e. 13 559–15 349 cal. yr BP) is called the 'Complexe téphrique CF1a/CF1b (Les Roches Tephra)'. The younger one called 'La Retombée de la Moutade' is dated at 11 360 $\pm$ 130 ^{14}C yr BP, i.e. 13 014–13 803 cal. yr BP. The stratigraphic position of the LAUT1 and LAUT2 show that they are older than the LST and, indeed, they may have been due to the Plinian eruptions of Le Puy de la Nugère (Chaîne des Puys, French Massif Central).

The most recent of the studies LST chronology propose an age of 12 880 cal. yr BP for the Laacher See eruption (Brauer *et al.*, 1999b; Zolitschka *et al.*, 2000). Taking into account the age of the Bølling–Older Dryas transition at 14 100 cal. yr BP (Amman *et al.*, 1994; Lowe *et al.*, 2001) and identified at 420 cm depth in core L6, the average sedimentation rate in core L6 is 0.6 $\pm$ 0.1 mm/yr during the first part of the Bølling. These two tie-points can be used to estimate an age for each tephra layer LAUT1 (4 cm from LST) and LAUT2 (8 cm from LST): 12 950 $\pm$ 10 cal. yr BP for LAUT1 and 13 020 $\pm$ 20 cal. yr BP for LAUT2.

Conclusion

The data presented reflect major transformations in terrestrial habitats and soil processes which may be related to climatic events characterising the Late Glacial warming and cooling phases. Over the interstadial/stadial cycle, changes in lithology are revealed by distinct shifts in the volume magnetic susceptibility measurements. Most of them can be interpreted as a direct response to climatic evolution: during warm intervals, the sediment lithology, κ and mineral content indicate increased lake productivity through carbonate precipitation and decreased delivery of detrital material. In contrast, cooler intervals show reduced lake productivity, catchment area instability and increased detrital input. These findings indicate the sensitivity of small carbonate lakes and their catchments to rapid climatic change.

Reconstruction of Late Glacial ecological changes shows important climate variability within the Bølling and Younger Dryas biozones. Several short interruptions in reforestation and soil stabilisation can be identified and linked with abrupt colder events occurring through the Bølling. The Younger Dryas is characterised by several steps in ecological and climate changes. At the beginning, an open landscape and detrital input attest to an abrupt cooling phase favouring aeolian erosion and transport. This is followed by a progressive improvement in climatic conditions towards the end of the period.

Overall, Lake Lautrey reveals two tephra layers, just before the LST which are most likely to be the product of volcanic eruptions in the 'Chaîne des Puys'. These tephra can be dated at Lake Le Lautrey around 13 ka cal. yr BP which is the most precise chronological information for volcanic activity of the 'Chaîne des Puys'. These two tephra occurrences constitute new tie-points for tephrochronological correlation of Jura and Northern Alps Late Glacial sedimentary records. Their characterisation appears necessary as they can be confused with the LST in low temporal resolution sequences. New detection is also necessary to determine their chronology and geographic extent. They constitute new chronostratigraphic markers for sequences where the LST is absent.

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Late-Glacial climatic changes in Eastern France (Lake Lautrey) from pollen, lake-levels, and chironomids

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Abstract

High-temporal resolution analyses of pollen, chironomid, and lake-level records from Lake Lautrey provide multi-proxy, quantitative estimates of climatic change during the Late-Glacial period in eastern France. Past temperature and moisture parameters were estimated using modern analogues and 'plant functional types' transfer-function methods for three pollen records obtained from different localities within the paleolake basin. The comparison of these methods shows that they provide generally similar climate signals, with the exception of the Bölling. Comparison of pollen- and chironomid-based temperature of the warmest month reconstructions generally agree, except during the Bölling. Major abrupt changes associated with the Oldest Dryas/Bölling, Alleröd/Younger Dryas, and the Younger Dryas/Preboreal transitions were quantified as well as other minor fluctuations related to the cold events (e.g., Preboreal oscillation). The temperature of the warmest month increased by ~5°C at the start of Bölling, and by 1.5°–3°C at the onset of the Holocene, while it fell by ca. 3° to 4°C at the beginning of Younger Dryas. The comparative analysis of the results based on the three Lautrey cores have highlighted significant differences in the climate reconstructions related to the location of each core, underlining the caution that is needed when studying single cores not taken from deepest part of lake basins.

Keywords: Chironomid; Climate; Pollen; Late-Glacial

Introduction

The study of palynological data and ice-cores has shown that the transition from the last glacial period to the present interglacial (ca. 14,000–9000 ^{14}C yr B.P.; 15,000–10,000 cal yr B.P.) was a period of special 'climatic' interest characterized by alternating cold and warm intervals with rapid transitions (Björck et al., 1998). These rapid and marked climate oscillations, associated with the successive steps of the deglaciation (Johnsen et al., 2001), have also been observed in Europe from various indicators such as pollen, macrofossils, oxygen isotopes, cladocera, beetles, and chironomids (Ammann et al., 2000; Birks and Ammann, 2000; Lemdahl, 2000; Lotter et al., 1992, 2000; Von Grafenstein et al., 2000; Walker et al., 2003). Quantitative estimates of paleotemperature have been inferred from most of these proxies, providing the basis for model/data comparisons (Renssen and Isarin, 2001; Renssen et al., 2001; Vandenberghe et al., 2001). Therefore, it is crucial that proxy-based reconstructions are

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realistic and reliable, and if possible, they should be validated by independent climate reconstructions from other proxies (Birks, 2003). In the continental (i.e., non-marine) realm, pollen, coleoptera, and chironomid records appear to offer the greatest potential to generate quantified climatic data. Pollen data offer the further advantage of giving information not only on temperature but also on precipitation because plant distributions respond to changes in summer warmth, winter cold, and moisture balance. However, at times of rapid climate changes such as the Late-Glacial period the vegetation development was influenced by other factors (such as migrational lags or edaphic conditions) that may induce biases in climatic reconstruction.

Recent studies show unambiguously a direct response of the Late-Glacial vegetation to rapid climate changes in Germany, Switzerland, and the Netherlands (Ammann et al., 2000; Birks and Ammann, 2000; Brauer et al., 1999; Hoek, 2001; Williams et al., 2002). In Switzerland, the warming that occurred during the Bølling/Allerød, and the Younger Dryas cold period have been particularly well documented by multi-proxy approaches (Ammann, 2000; Ammann et al., 1994; Birks and Ammann, 2000; Lotter et al., 1992, 2000; Walker, 2001). In the French Jura, located close to the Swiss Plateau, the Late-Glacial vegetational changes characterized by a transition from an open vegetation with steppe elements to a more or less forested landscape have been well established from numerous palynological and lake records (Bégeot et al., 2000; Magny, 2001; Magny and Ruffaldi, 1995; Richard and Bégeot, 2000).

However, quantitative estimates of the Late-Glacial climate from these records are still rare. In the Swiss Jura and the French pre-Alps, high-resolution pollen and lake-level records from le Locle and St-Jorioz have been used to precisely reconstruct the climate of the Younger Dryas and early Holocene key events such as the Preboreal oscillation and the 8200 yr event (Magny et al., 2001, 2003). However, no reliable quantified climatic data that span the entire Late-Glacial are available for the Jura Mountains.

The purpose of this study is to obtain robust and precise quantitative estimates of the Late-Glacial climate of eastern France from high-resolution pollen and lake-level records taken from Lake Lautrey, Jura Mountains. We attempt to estimate the magnitude of temperature and precipitation changes at the onset and end of the Oldest and Younger Dryas, as well as during the known minor Late-Glacial oscillations at ca. 13,900, 13,000, and 11,200 cal yr B.P. This study is based on a comparative analysis of pollen data from three cores taken at different locations in the basin of the Lake Lautrey. Our objective is to examine the three palynological records, which cover the whole Late-Glacial, in particular considering the paleovegetation records that they provide, and their paleoclimatic interpretation. Examination of the reconstructions from different cores allows greater insight into the prevailing climatic conditions than would the results from any of the cores alone. Huntley et al. (1999) have shown that at Lago Grande di Monticchio (Italy), paleoclimate reconstructions from three pollen sequences taken at different locations exhibit apparent discrepancies in the reconstruction of the Late-Glacial climate and that the most accurate reconstruction was obtained from the mid-lake pollen record, which better represents the regional vegetation.

Furthermore, the paleoclimatic reconstructions presented here are based on a new modern pollen dataset. Samples from cold steppes and desert of the Tibetan plateau and pioneer vegetation from the Scandes Mountains have been compiled in order to limit the lack of 'good modern pollen analogues' for Late-Glacial key periods, such as the end of the Oldest Dryas and the onset of the Bølling. A comparative analysis of the inferred climate provided by two quantitative methods is proposed here. Such approaches have been successfully tested for other key periods and enable us to provide more precise and robust climate estimates than those based on only one method (Klotz et al., 2003; Lotter et al., 2000; Peyron et al., 2000). Here, the "modern analogue technique" (Guiot, 1990), and a transfer function based on "plant functional types," which is particularly efficient in no-analogue situations (Peyron et al., 1998), are used. Furthermore, the pollen-inferred summer temperature estimates will be compared with independent paleotemperatures inferred from chironomid records (Heiri and Millet, 2005).

Environmental setting

Lake Lautrey (46°35′N, 5°51E) is a small residual lake located at 788 m a.s.l. in the Jura Mountains, eastern France (Fig. 1). The climate of the Jura Mountains is semi-continental with oceanic influences. At Lake Lautrey, the mean annual temperature is 8°C, the mean temperature of the warmest month is 16°C, and annual precipitation ranges from 1650 to 1750 mm. The modern catchment area is dominated by a dense mixed forest of coniferous and deciduous trees.

Based on a detailed geophysical exploration of the lake deposits (Bossuet et al., 2000), sediment cores were taken at three different places in the former lake basin. Core 6, which was subsequently used for the multi-proxy study, is located ca. 80 m north of the present-day lake. Core 105 was extracted from the deeper part ca. 80 m northwest of the lake, and core 375 ca. 200 m southwest of the lake (Fig. 1).

The chronology of core 6 is based on twelve AMS radiocarbon dates of terrestrial macrofossils (Table 1). An age-depth model (Fig. 2) was developed for the core 6 by M. Magny (personal communication), based on these dates and on isotopic measurements of lacustrine carbonates

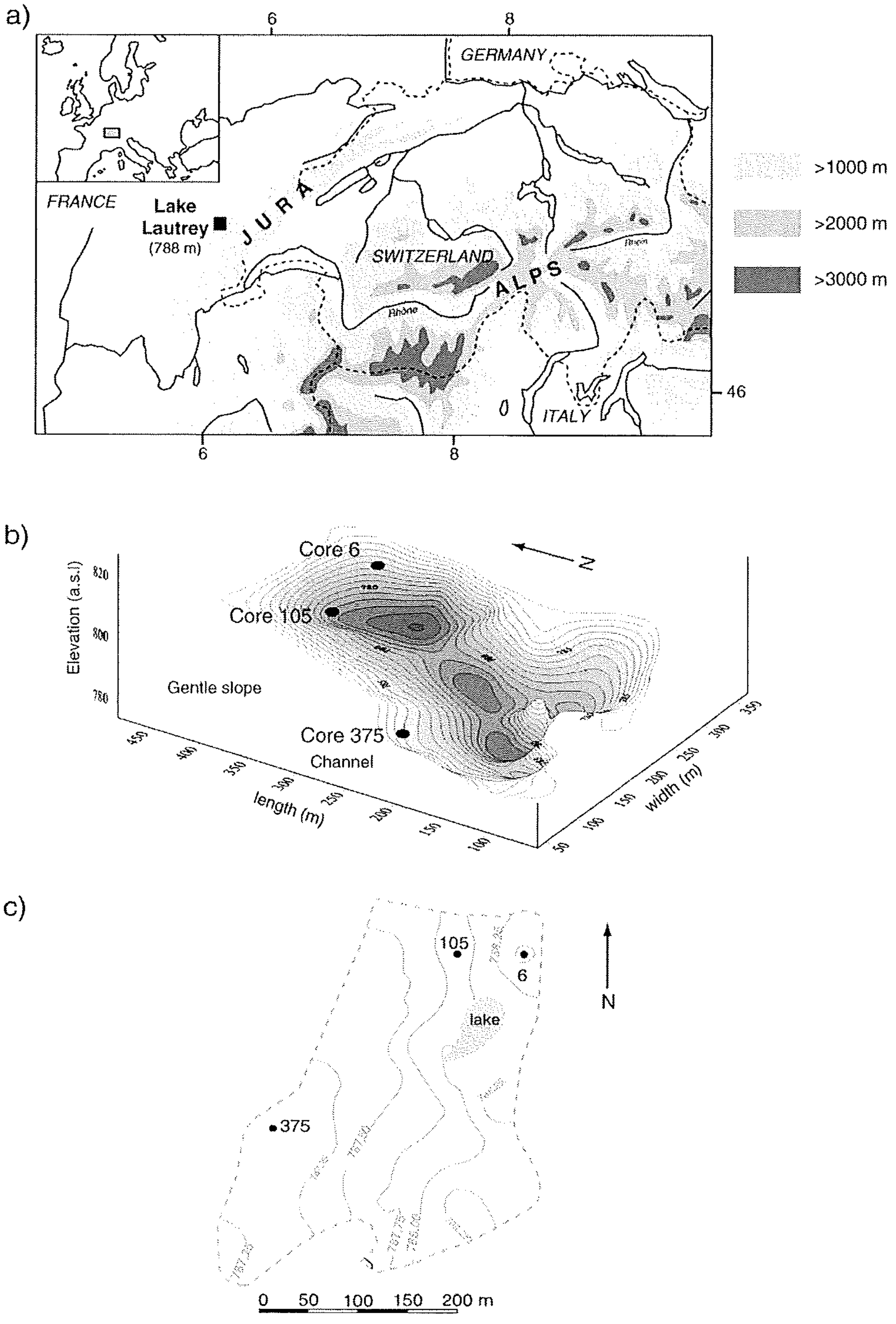

Figure 1. (a) Location of Lake Lautrey, (b) Lake Lautrey basin morphometry showing the substratum before Late-Glacial and Holocene sediments filling, and (c) location of the three coring sites; dashed line indicates the limit of the lake during the Late-Glacial period; shaded area shows the present-day residual Lake Lautrey (from Bossuet et al., 2000).

correlated to the Greenland Ice Core Project (GRIP) $\delta^{18}O$ record. Further, the Laacher See Tephra (LST), dated to 12,880 cal yr B.P., was found in the three sequences and was used as a time marker. Ages of the sediment sections between the time markers were determined by linear interpolation (Fig. 2).

Pollen data

Fossil sequences

The study of the Late-Glacial vegetation development at Lake Lautrey is based on the results of the three pollen

Table 1
Lautrey (core 6) radiocarbon dates along with calibrated ages, laboratory number, and material dated

Depth (cm)	Age, ^{14}C yr B.P.	Age, cal yr B.P. (2 sigmas)	Laboratory number (Vienne)	Material
286	9975 ± 45	11,904–11,130	VERA-1717	Twigs
262	10,120 ± 40	12,269–11,358	VERA-1722	Twigs
296	10,000 ± 40	11,918–11,257	VERA-1716	Twigs
305	10,140 ± 35	12,282–11,442	VERA-1715	Twigs
324	10,960 ± 45	13,154–12,657	VERA-1724	Twigs
326	11,100 ± 45	14,191–12,683	VERA-1725	Twigs
366	11,575 ± 45	13,836–13,317	VERA-1726	Twigs
384	11,695 ± 45	15,092–13,442	VERA-1727	Twigs
398	11,825 ± 45	15,205–13,587	VERA-1728	Twigs
416	12,170 ± 60	15,400–13,836	POZ-4496	Twigs
426	12,430 ± 45	15,517–14,139	VERA-1729	Twigs
445	12,590 ± 45	15,592–14,223	VERA-1730	Twigs

sequences (P. Ruffaldi and C. Bégeot, unpublished data). Biostratigraphical correlation of the pollen diagrams was made using common trends in the pollen curves of the main taxa. The base of the cores, as illustrated by core 6 (Fig. 3), is dominated by high percentages of herbaceous pollen, mainly *Artemisia*, *Poaceae* and other heliophilous taxa (e.g., *Chenopodiaceae, Helianthemum*). These data indicate an open herbaceous landscape and can be correlated with the Oldest Dryas biozone. Low *Pinus* values (less than 20%) occur as a result of long distance transport. The Bölling biozone (Ammann and Lotter, 1989; Ammann et al., 1996) is characterized by an increase in *Juniperus* pollen percentages followed by a decline towards the upper boundary of the zone (Fig. 3). The expansion of shrubby vegetation is also suggested by the presence of *Salix* and *Hippophaë*. The second part of this period corresponds to the establishment of *Betula* woodland. The *Betula* pollen curve shows successive depressions usually combined with higher percentages of *Juniperus* and/or NAP. These changes may reflect regressive phases in the vegetation development previously observed in pollen diagrams from the Jura during this period (Bégeot et al., 2000). During the Allerød biozone, the regional immigration of pine is indicated by a gradual rise of *Pinus* pollen percentages. The presence of macrofossils of *Betula pendula* combined with high pollen

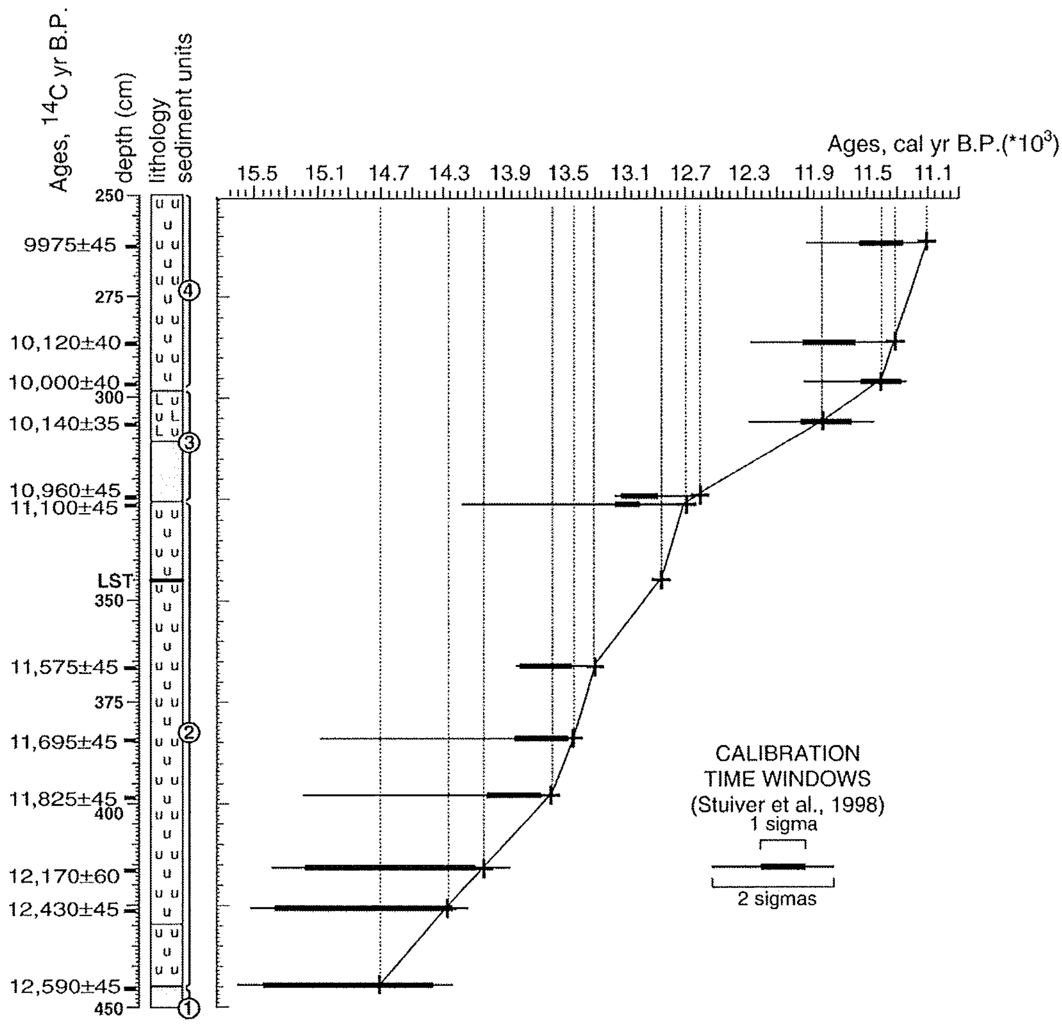

Figure 2. Age-depth model developed for the core 6.

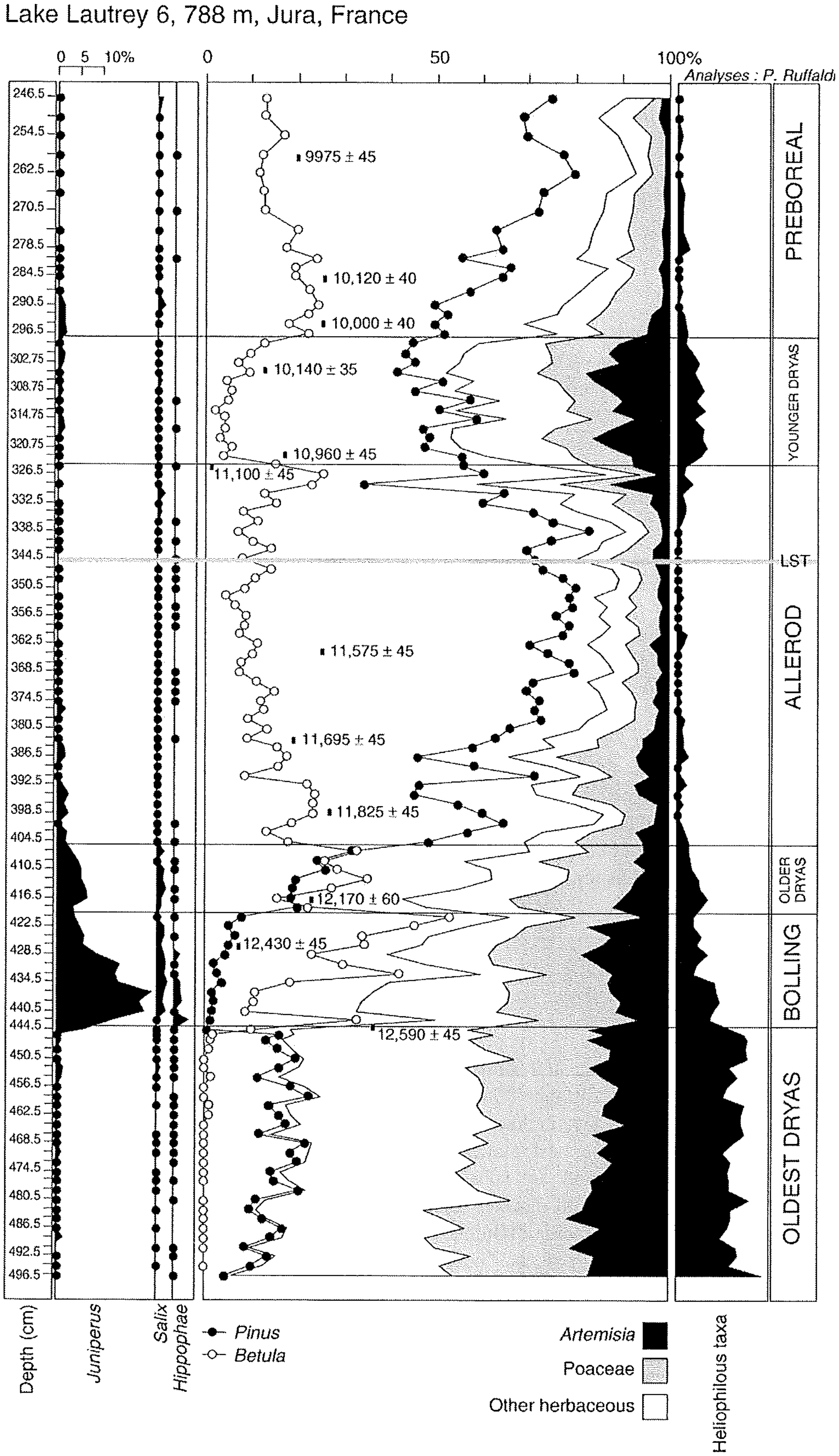

Figure 3. Simplified pollen diagram of Lake Lautrey core 6, showing location of dates ^{14}C yr B.P. and biozones.

values show that this taxon remains a significant element in the vegetation. Just before the LST, a strong short rise in *Betula* values of core 105 (P. Ruffaldi, unpublished data) may be correlated to the Gerzensee oscillation (Lotter et al., 1992). During the Younger Dryas biozone, the other cores show high percentages of arboreal pollen (Fig. 3) probably due to their proximity to the shore (Fig. 1), while the core 105 displays a major rise in herbaceous taxa. The Younger Dryas/Preboreal transition is marked by the rapid colonization of near-shore areas by dense *Pinus–Betula* forests.

Modern surface samples

The modern dataset used in previous climatic reconstructions included pollen surface samples mainly located in Europe, Kazakhstan, and the Mediterranean area (Peyron et al., 1998). To improve the reconstruction of the Late-Glacial climate, in particular the estimates of the temperature of the warmest month, we have updated the dataset by adding 118 samples from the Tibetan plateau cold steppes (Van Campo et al., 1996; Yu et al., 2001), 37 from the Scandes Mountains, and 17 from Spain (Sanchez Goni and Hannon, 1999). The updated dataset now contains 868 surface samples.

The climatic parameters that play a determinant role on the distribution of the vegetation and the related pollen assemblages have been interpolated at each modern pollen site from meteorological stations by using an Artificial Neural Networks (ANN) technique (Peyron et al., 1998). ANN are computer systems able to predict new observations from other observations after executing a process of "learning" from existing data. The parameters selected are the mean temperature of the warmest month, the annual precipitation, and the ratio of real to potential evapotranspiration.

Methods

Modern analogues technique (MAT)

The similarity between each fossil sample and modern pollen assemblage is evaluated by a chord distance (Guiot, 1990). Usually, a selection of 5–10 modern spectra that have the smallest distance are considered as the best modern analogues of the given pollen spectrum and are subsequently used for the reconstruction. If the chord distance is above a threshold defined by a Monte-Carlo method (Guiot, 1990), the modern samples are considered as "bad analogues", and are not taken into account in the reconstruction. Estimates of climatic parameters are obtained by taking a weighted average of the values for all selected best modern analogues, where the weights used are the inverse of the chord distance. To reduce uncertainties (caused, for example, by a lack of perfect modern analogues) in the reconstruction, two additional constraints have been used. The first one is a constraint by biome, in which a biome is assigned to modern and fossil pollen assemblages, using the method of Prentice et al. (1996). The biomes assigned to the selected modern analogues are compared to the biome assigned to the fossil assemblage, and only the analogues with consistent biomes are retained for the analogue matching step. The second constraint is based on lake-level fluctuations (Guiot et al., 1993a). Analogues that were not compatible with the Lake Lautrey lake-level reconstruction (M. Magny, personal communication) were rejected (Magny et al., 2001). The number of best analogues selected is constant through the whole core (core 6: 10 analogues because of the lake-level constraint; cores 105 and 375: 5 analogues).

Plant functional type method (PFT)

The main principle and different steps of this method are fully described in Peyron et al. (1998, 2000). Plant functional types are broad classes of plants defined by their structural and functional features (Prentice et al., 1996). The PFT method is a transfer function in which the pollen percentages are first transformed into PFT scores (the square root of pollen %; threshold 0.5%). PFT scores derived from the modern pollen samples are then calibrated in terms of climatic parameters. In this procedure, the calibration between the scores of PFTs and climate parameters is achieved by non-linear regression using an ANN (Peyron et al., 1998). The modern pollen original dataset has been divided into training and verification sets. The ANN is calibrated on samples from the training set and samples of the verification set are used to verify the prediction of the climate parameters. The network has been calibrated by randomly extracting, with replacement, 868 samples from the modern data-set and then verified on the unused samples. Following a bootstrap procedure, this is repeated 30 times to obtain a statistically valid estimate of the error rates. The calibration is then applied to the fossil pollen scores to infer the paleoclimate estimates.

The two methods (MAT and PFT) have been used to reconstruct the mean temperature of the warmest month (T_w), total annual precipitation (P_{ann}), and the ratio of real to potential evapotranspiration (E/PE).

Reliability of the reconstructions

To evaluate the reliability of both methods, the climate parameters for each surface sample were estimated using the other modern samples. The difference between present-day climate data at the pollen sites and the estimated climate at each site indicates the quality of the method. As both methods have high coefficients of correlation between the observed and estimated parameters (Table 2), both methods appear to be reliable for the interpretation of paleoclimate change. As expected, the correlation between observations and reconstructions is excellent for the

Table 2
Correlation coefficients between observed and reconstructed values of climate parameters obtained from application of both MAT and PFT approaches to the modern pollen samples

Climatic parameter	MAT		PFT		
	Correlation coefficient	RMSE	Correlation coefficient for the calibration	Correlation coefficient for the verification	RMSE
Temperature of the warmest month (°C)	0.94	2.0	0.88 ± 0.016	0.87 ± 0.020	3.1
Annual precipitation (mm)	0.90	162	0.81 ± 0.014	0.80 ± 0.030	212
Ratio of real to potential evaporation (%)	0.94	9.8	0.91 ± 0.007	0.90 ± 0.012	13.3

For the PFT approach, coefficients of correlation and standard deviation after the calibration and the verification procedures are indicated.

temperature variables but lower for the hydrological parameters. The best estimated variable related to vegetation water-stress is E/PE, which has a correlation comparable to the temperature. These validation tests may suggest that the MAT reconstructions are more accurate than the PFT method (Table 2). In theory, the artificial neural networks used in the calibration are able to learn and reproduce any pattern perfectly. However, if the network is fitted too closely to the input data, there is a risk that estimates of new data will be incorrect. Consequently, even if the correlation coefficients are lower with the PFT method than with the MAT, we consider the calibration procedure to be optimal because the relationships between the input and the output variables is learnt in a generalized fashion and can therefore be used for predictions of unknown outputs.

Results/discussion

The paleoclimatic reconstructions at Lake Lautrey show that the climatic estimates derived from the MAT and the PFT method for core 6 are remarkably similar from 16,000 to 10,500 cal yr B.P. (Fig. 4). Statistically significant deviations are only evident during the Bölling for E/PE and during the Bölling and the Older Dryas for P_{ann}. Although plotted deviations between the reconstructions may sometimes appear to be high, for instance annual

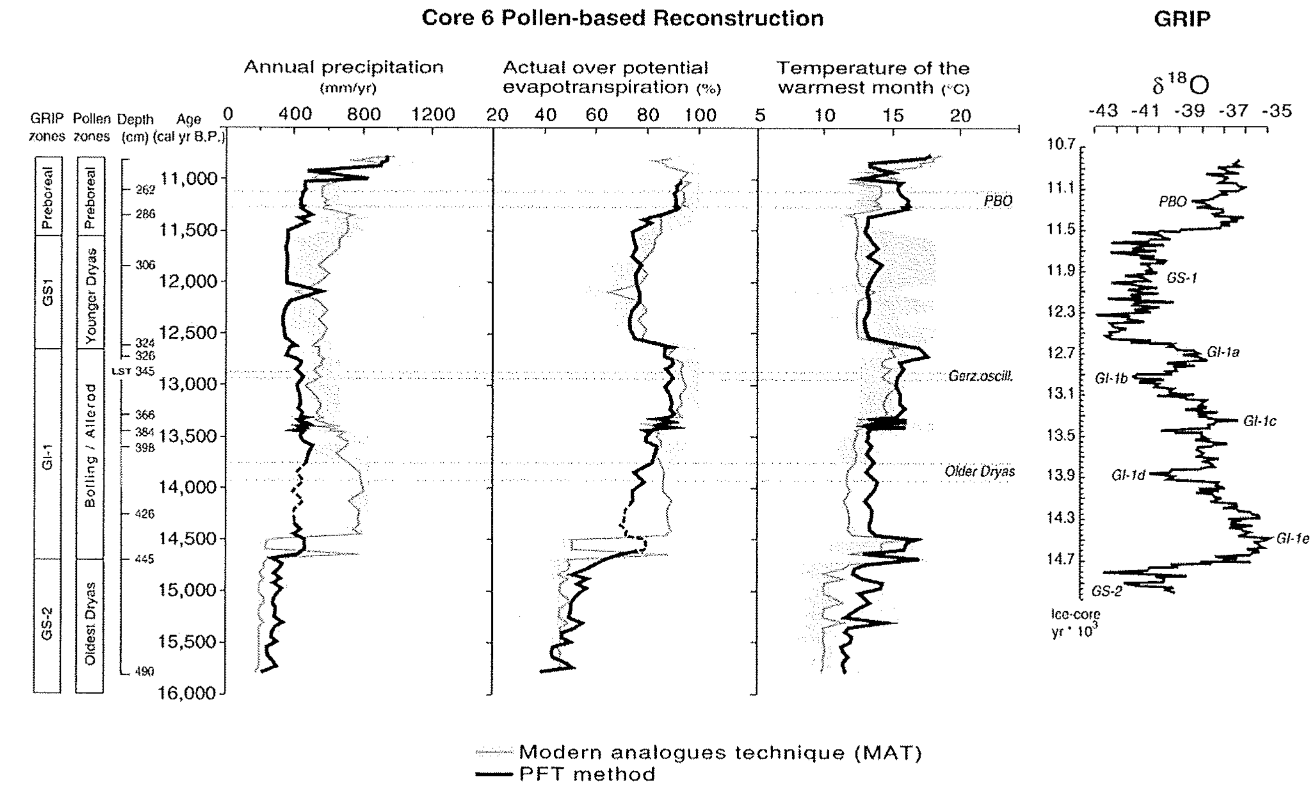

Figure 4. Pollen-inferred climatic estimates at Lake Lautrey from core 6 using the Modern Analogues Technique (dark grey), and Plant Functional Types (black). For clarity, only the error bars calculated with the MAT are plotted (grey interval). Dashed lines in the PFT-based curve show where significant deviations between the two methods are observed for each reconstructed climate parameters. Dotted lines (grey) delimit the boundaries of the climatic oscillations: Older Dryas, Gerzensee oscillation, Preboreal oscillation (PBO). LST is the abbreviation for Laacher See Tephra. GRIP oxygen isotope ratio is indicated (Johnsen et al., 2001). GS-2 represents the last cold phase of the Pleniglacial, GI-1 the Late-Glacial Interstadial, and GS-1 the Younger Dryas cold event.

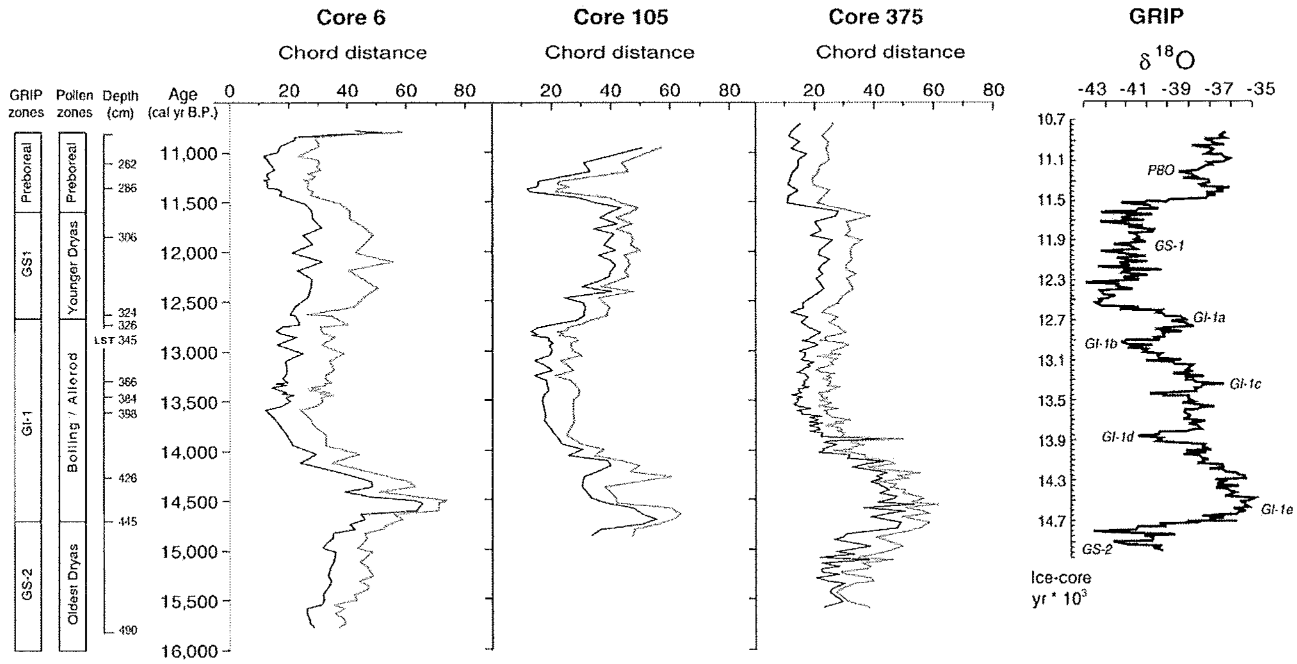

Figure 5. Similarity index (Chord distance) measuring the quality of the modern analogues for cores 6, 105, and 375, Lake Lautrey. The figure shows the calculated distance of the first and the last of the best analogues selected.

precipitation during the Younger Dryas, the two reconstructions are similar. The deviations are more pronounced for P_{ann} than for E/PE and T_w, which may be due to the overall smaller range of these latter variables. To provide an independent validation (Fig. 5), the pollen-based temperature of the warmest month estimates have been compared with a July air temperature curve inferred from the chironomid records (Heiri and Millet, 2005), using weighted

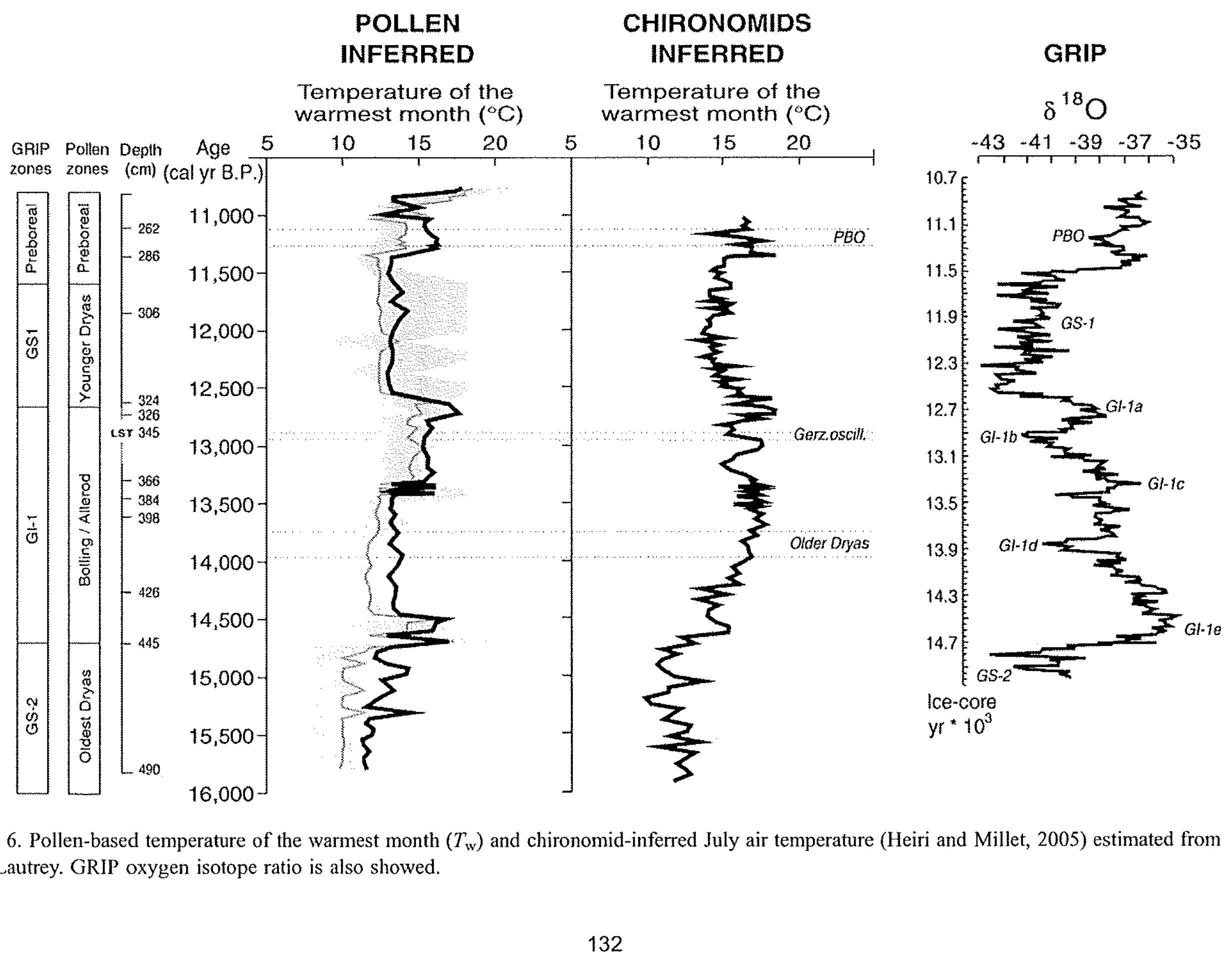

Figure 6. Pollen-based temperature of the warmest month (T_w) and chironomid-inferred July air temperature (Heiri and Millet, 2005) estimated from core 6, Lake Lautrey. GRIP oxygen isotope ratio is also showed.

averaging-partial least squares regression (Heiri et al., 2003).

By using the results from the two methods, the chironomid-inferred reconstruction, and the oxygen isotope ratios from the Greenland ice records as an additional paleoclimate reference (Figs. 4 and 6), it is possible to establish a general scheme of climatic variations at Lake Lautrey as follows:

Oldest Dryas, before 14,700 cal yr B.P.

At Lake Lautrey, all cores (Figs. 4–7) suggest that before 14,700 cal yr B.P., T_w was lower than today, with values fluctuating around 10°–12.5°C (3°–5.5°C less than today). The chironomid-inferred July temperatures oscillate around 11°–12°C and agree particularly well with the pollen inferred T_w (Fig. 6). Further, the inferred precipitation at Lake Lautrey indicates that climatic conditions were much drier than today. During the Oldest Dryas, E/PE reconstructed values oscillate around 50%, and P_{ann} estimates are low, about 200 to 300 mm. In general, it appears that the climatic reconstructions obtained from the 3 cores are remarkably similar for both the temperature and the hydrological parameters, while the estimates obtained using the MAT agree well with those of the PFT method.

Some problems arise, however, when our reconstruction is compared with quantifications inferred from other proxies. The few existing climate reconstructions concerning the Oldest Dryas (Table 3) indicate slightly cooler temperatures than in our reconstruction. In Switzerland, fossil coleopteran assemblages provide quantitative estimates of summer paleotemperatures ranging from 9° to 10°C (Coope and Elias, 2000; Gaillard and Lemdahl, 1994). Renssen and Isarin (2001) have reconstructed the July air temperature in France and west-central Europe from numerous botanical data. Based on the method of "climate indicator plant species", they have reconstructed minimum July temperatures ranging from 13° to 15°C at sea-level (i.e., 9° to 10.5°C at the altitude of Lake Lautrey). Further, they have simulated the Late-Glacial climate with the Atmospheric General Circulation Model ECHAM4, and obtained July temperatures during GS-2a ranging from 10.5° to 15.5°C at 800 m a.s.l. We can conclude that both the model simulation and data reconstructions done by Renssen and Isarin (2001) agree well with our estimates during GS-2a, even if the model suggests slightly higher

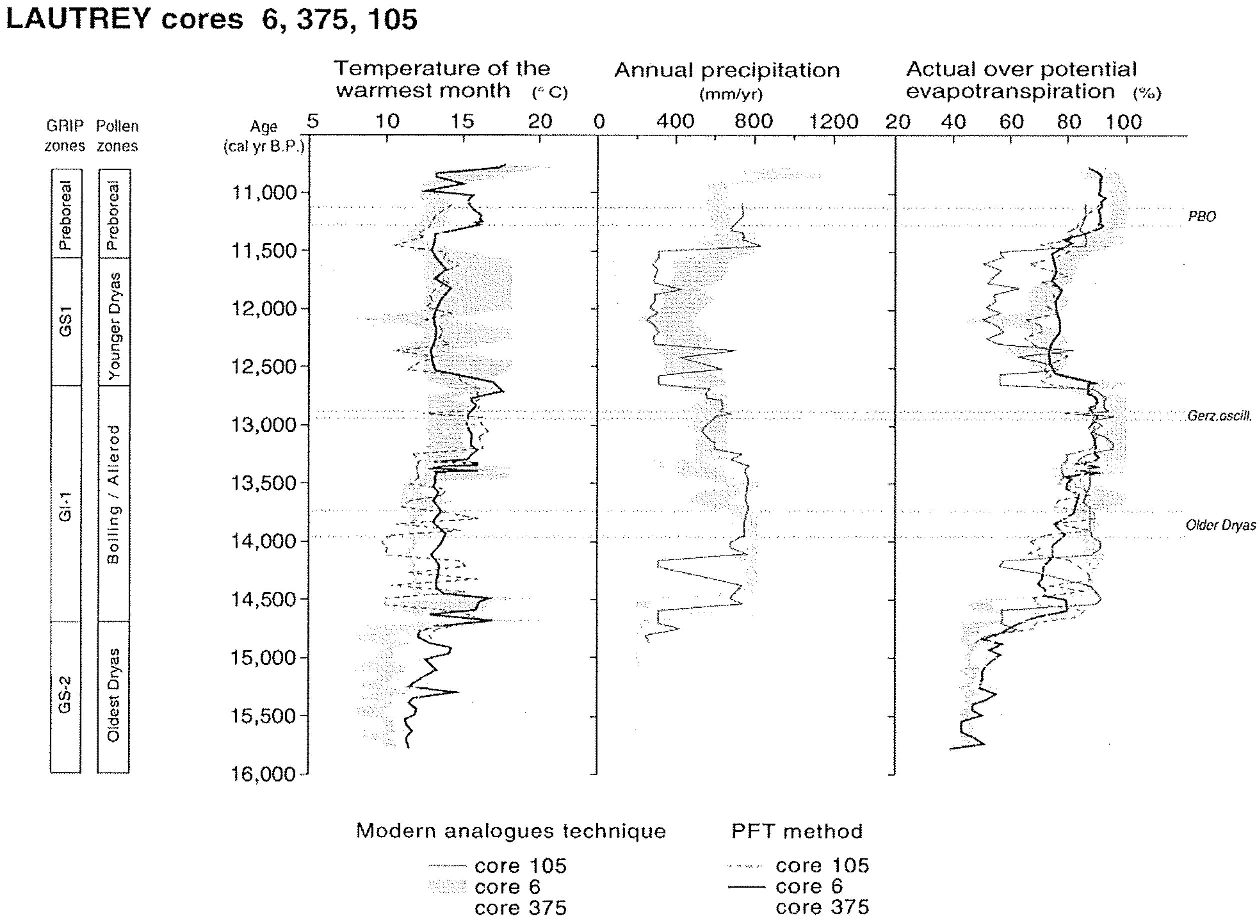

Figure 7. Pollen-inferred climatic parameters estimated at Lake Lautrey from cores 6, 105, and 375 using the MAT and PFT methods. For clarity, only the error bars of core 6 calculated with the MAT are plotted (grey interval).

Table 3
Comparison of Late-Glacial quantitative reconstructions of summer temperature in west central Europe using various proxies and approaches (MAT: modern analogues technique, PFT: plant functional type method, IF: inference model, MCR: Mutual Climatic Range analysis)

Sites or regions	References	Data/method	Oldest Dryas	Bölling	Gerzensee oscillation	Alleröd/Y. Dryas transition	Younger Dryas second part	Y. Dryas/Preboreal transition	Preboreal oscillation
Lake Lautrey, 788 m a.s.l.	This study	Chironomids	11° to 12°C	15.5° to 16°C (+5°C)	16.8° to 15°C (−2°C)	18.4° to 15.5°C (−3°C)	13.6° to 13.9°C (+0.3°C)	13.9° to 15.5°C (+1.5°C)	17.5° to 14°C (−3.5°C)
		Pollen data (PFT)	11.5° to 12°C	17°C (+5°C)		17° to 1.2°C (−4°C)		lag. 13° to 16°C (+3°C)	16.2° to 15.2°C (−1°C)
		Pollen data (MAT)	10.5° to 11.5°C	18°C (+6°C)		17° to 12.8°C (−4°C)			
Lake Le Locle, 915 m a.s.l.	Magny et al., 2001	Pollen and lake–level data (MAT)					13.2° to 14.2°C (+1°C)	14° to 16°C (+2°C)	16° to 14°C (−2°C)
Gerzensee, 603 m a.s.l.	Lotter et al., 2000	Pollen data (MAT)			12° to 10.7°C (−1.3°C)	11.5° to 8.7°C (−3°C)	9° to 10°C (+1°C)	10° to 12°C (+2°C)	12° to 11.5°C (−0.5°C)
		Cladocera (IM)			12° to 11.2°C (−0.8°C)			9° to 11.5°C (+2.5°C)	
Hauterive–Champréveyre, 428 m a.s.l. (Lake Neuchatel)	Coope and Elias, 2000	Coleoptera (MCR)	9°C	16°C					
Grand–Marais, 587 m a.s.l.	Gaillard and Lemdahl, 1994	Coleoptera (MCR)	10°C	15°C					
South–Western Switzerland	Wohlfarth et al., 1994	Pollen and isotope data		12° to 14°C		−3° to 4°C			
West–Central Europe	Isarin and Bohncke, 1999;	Geological–ecological data	9° to 10.5°C	12.5° to 15.5°C (+3.5+5°C)			10° to 12°C	13° to 15°C (+3°C)	
	Renssen and Isarin, 2001	AGCM	10.5° to 15.5°C	15.5° to 20.5°C (+5°C)			16°C	21°C (+5°C)	

temperatures. Our results also concur with the CCM1 model which simulates a reduction in summer temperatures by 5.5°C compared to today (Kutzbach et al., 1998).

Oldest Dryas-Bølling transition, ~14,700 cal yr B.P.

At ~14,700 cal yr B.P., the end of the Late Pleniglacial is marked by a strong warming (Figs. 4–7). Pollen-inferred T_w increase by 5°C and reach values comparable to modern temperature (~16°C). At the same time, the chironomid-inferred July temperature abruptly increases by 4.5°C, rising to 15.5°C in the earliest part of the Bølling (Fig. 6). These reconstructions are remarkably similar. In west-central Europe, Renssen and Isarin (2001) have reconstructed an increase in July temperatures of 3° to 4°C (1° to 2°C less than our data) at the Oldest Dryas-Bølling transition. Using the ECHAM model, they simulated a temperature increase of 5°C (Table 3). It is of note that their simulated Bölling July temperatures are about 5°C higher than their reconstructed values. Given the fact that the pollen-based estimates by the 'climate indicator plant species' method are minimal values and that the model tends to produce too warm conditions, Renssen and Isarin (2001) propose that the 'true' Bölling temperature was intermediate between the reconstructed and simulated values (i.e., between 12° to 15°C and 15° to 20°C). This 'true' temperature is closer to our reconstruction and also to the beetle-inferred paleotemperatures. In Switzerland, beetle assemblages show an increase in summer temperatures that is similar to our values (i.e., 5° to 6°C), and reached values comparable to the modern ones (Coope and Elias, 2000; Gaillard and Lemdahl, 1994). In addition, a minimum mean July temperature between 14° and 16°C has been reconstructed in central-west Germany (Bos, 2001), in good agreement with our results.

In our reconstruction, P_{ann} and E/PE values also increase abruptly to reach 800 mm, and 78–90%, respectively. Again, the climatic reconstructions obtained from the three cores are remarkably similar for both the temperature and the hydrological parameters (Fig. 7). Further, both approaches provide similar temperature reconstructions and a similar increase in P_{ann}.

Bølling/Allerød, ~14,700 to ~ 12,650 cal yr B.P.

From ~14,700 to ~14,000 cal yr B.P., the pollen-based reconstructions suggest a surprising decrease in T_w, in contrast to the GRIP $\delta^{18}O$ and to the chironomid-based temperature which gradually increases in two steps, from 14° to 16.5°C (Fig. 6). In central-west Germany, paleotemperatures reconstructed during this period are also slightly lower than during the Oldest Dryas-Bølling transition (Bos, 2001), in agreement with our data. However, our results should be interpreted with caution because the first part of Bølling is marked by more pronounced discrepancies between the individual reconstructions from the three cores and by significant deviations between the MAT and the PFT methods (Figs. 4 and 7). The differences between the three cores are particularly pronounced with respect to T_w and E/PE. T_w reconstructed from core 375 is around 10°C during the Bølling, whereas T_w reconstructed from core 105 oscillates widely around 15°C. This latter value is in better agreement with the beetle-inferred estimates from Switzerland (Coope and Elias, 2000; Gaillard and Lemdahl, 1994) and with the temperatures simulated by the CCM1 model (Kutzbach et al., 1998). These differences may be related to the location of each core, which represent different pollen depositional environments. Core 105 was collected in the deepest part of Lake Lautrey and probably reflects a more regional picture of vegetation whereas core 375, closer to the shore, reflects principally the local vegetation within the basin. At Lago Grande di Monticchio (Italy), paleoclimate reconstructions from three pollen sequences taken at different locations also exhibit apparent discrepancies in the reconstructions of the Bølling climate (Huntley et al., 1999).

Significant deviations between P_{ann} and E/PE reconstructed from both methods occur only in the first part of Bølling (Fig. 4). The results obtained using the modern analogues technique are characterized by a deterioration in the quality of the analogues (an increase in chord distance) as well as by numerous 'no-analogue' situations (Fig. 5). These deviations generally correspond to features of the pollen assemblages that are interpreted in different ways by the two methods. At Lake Lautrey, the pollen assemblages that characterize the Bølling are dominated by pioneer taxa such as *Juniperus*, *Betula*, and *Salix*. These taxa are present in the modern pollen spectra from the Scandes Mountains in our modern database, but at much lower percentages than in the fossil samples. Additional modern samples are required from this area to improve the reconstruction of the complex climatic conditions occurring during this period.

During the Allerød at Lake Lautrey, the temperatures inferred from pollen and chironomids follow a similar pattern, with an increase to and fluctuations around present-day values. This reconstruction agrees well with the temperature reconstructed from coleopteran assemblages at Gersenzee and Grand-Marais in Switzerland (Gaillard and Lemdahl, 1994; Lemdahl, 2000). However, the agreement is less good with reconstructions based on fossil cladoceran and pollen assemblages from Gerzensee, which showed July temperatures between 3.5° and 5.5°C cooler than the present (Lotter et al., 2000). At Lake Lautrey, the climatic conditions remained humid during most of the Allerød, with E/PE reaching 80 to 90%, but precipitation beginning to decrease from 800 mm at ~13,700 cal yr B.P. to ~500 mm at ~12,650 cal yr B.P.

Younger Dryas, ~12,650 to ~11,500 cal yr B.P.

The transition to the Younger Dryas is characterized by a rapid decrease in temperature and E/PE (Figs. 4–7). T_w

values from the three cores range between 12 and 14°C (i.e., ~3° to 4°C less than GI-1). E/PE values vary from 65 to 75%. In the chironomid-inferred reconstruction, the cooling, while also rapid, is more gradual, with July air temperatures oscillating slightly around 13.5°C (Fig. 6). The Younger Dryas is particularly well represented in core 105 with a sampling resolution of about 25 years (P. Ruffaldi, unpublished data). Core 105 displays a major rise in pollen herbaceous taxa (e.g., *Artemisia*, *Chenopodiaceae*), and a return of *Juniperus*. The period may be split into two parts. In the first part, percentages of *Pinus* pollen remain high whereas *Betula* pollen declines. The second part is characterized by a slight recurrence in *Betula* pollen and a fall in *Pinus* pollen. The climatic reconstruction from core 105 suggests colder conditions for the first part of the Younger Dryas than for the second part (Bohncke et al., 1993). The reconstruction of P_{ann} from core 105 suggests a more pronounced decrease, with values of ~300–350 mm, than the two other cores (Fig. 7). The differences are caused by the fact that the Younger Dryas event is more strongly represented in core 105, with a higher proportion of herbaceous taxa than in the other two cores. Cores 6 and 375 show higher percentages of *Pinus* pollen probably as a result of a concentration of this pollen near the shore: higher amounts of vesiculate pollen types (e.g., *Pinus*) in proximity to the shore is a phenomenon often observed in palynological studies and is assumed to be the effect of longer flotation time of these saccate pollen grains (Ammann, 1994). The discrepancies observed in the climatic reconstructions may reflect the fact that pollen data may sometimes give a biased picture of the vegetation (Birks, 2003). The high arboreal percentages in cores 6 and 375 may indicate that the Younger Dryas climatic deterioration had a less marked impact upon the local vegetation than the regional vegetation, with greater tree cover persisting at the lake. In contrast, the pollen records from core 105 indicate a large reduction of the forested surface over a large portion of the jurassian massif, which corresponds to a depression of the treeline to at least 800 m a.s.l. (Bégeot et al., 2000). The absence of arboreal macrofossils and the marked decrease in pollen concentrations further support the opening of the forest around the lake. This is equally shown in the sediment and chironomid records (M. Magny, personal communication; Millet et al., 2003). Core 105 was collected in the deepest part of the lake and it is reasonable to expect that it reflects a more regional vegetation record.

Only the T_w curve from core 105 (Fig. 7) shows that the first part of the Younger Dryas (~11°C) was colder than the second part (13° to 14°C). This pattern is in agreement with results obtained from the Swiss Jura, Switzerland, Germany, and central and northwestern Europe (Table 3; Birks and Ammann, 2000; Bos, 2001; Isarin and Bohncke, 1999; Magny et al., 2001). At Le Locle, at 915 m a.s.l. on the Jura Mountains, the first part of the Younger Dryas is characterized by lower estimates of T_w (13° to 13.5°C) than the second part (13.5° to 15°C). The minimum mean July temperature reconstructed by Bos (2001) in central-west Germany (~11.5°C during the earlier Younger Dryas and ~13°C during the later Younger Dryas) agrees particularly well with our data. The same trend is shown at Gerzensee on the Swiss plateau. At this site, mean summer temperatures reconstructed from pollen, coleopteran, and cladocera data also show a slight rise over the Younger Dryas from 9° to 10°C (Birks and Ammann, 2000; Lemdahl, 2000). However, the temperatures reconstructed at Gerzensee at the end of the GS-1 appear cooler, by 3° to 4°C, than the values estimated here. The differences in the temperature values estimated at Gerzensee and at Lake Lautrey are probably related to differences in the modern pollen databases used and to the methods involved in the paleoclimate reconstructions.

In west-central Europe, Renssen and Isarin (2001) reconstructed a July temperature close to 11.5°C at the end of the Younger Dryas from paleobotanical data. Their ECHAM model simulation produced a mean July temperature of ~15.3°C for the Lake Lautrey altitude. Given that the uncertainty in their pollen-based reconstruction is about 1° to 2°C (Isarin and Bohncke, 1999), we suggest that their results agree with the reconstruction presented here. As before, the model temperatures are higher than the reconstructed temperatures, suggesting that the "true" Younger Dryas temperature was probably intermediate between the reconstructed and simulated values.

Younger Dryas/Preboreal transition, ~11,500 cal yr B.P.

The transition is characterized at Lake Lautrey by a rise of ~2.5° to 3°C in the temperature estimates based on the pollen and chironomid records (Figs. 4–7). This value is close to that reconstructed from pollen data at le Locle (Magny et al., 2001), Gerzensee (Birks and Ammann, 2000; Lotter et al., 2000), and central-west Germany (Bos, 2001). The results also support the July temperature increase of 3°C in Central Europe (Renssen and Isarin, 2001) during the Younger Dryas-Holocene transition. However, the ECHAM model produces T_w that are again slightly higher than the reconstructions.

Late-Glacial oscillations

The GRIP $\delta^{18}O$ curve shows evidence of three climate oscillations during the GI-1 stadial (Fig. 4; Björck et al., 1998; Johnsen et al., 2001). The paleotemperature record as reconstructed by the PFT method at Lake Lautrey shows a first slight cooling at around 13,900 cal yr B.P. which is synchronous to the GI-1d event (Fig. 4). At this time, both the pollen and chironomid records indicate a weak decrease of temperature by 0.75–1°C and a decrease of E/PE by 5 to 10% (Figs. 6 and 7). The decline in T_w is much more pronounced in the core 105 reconstruction (Fig. 7). This event corresponds to the Older Dryas cold event and to the Aegelsee oscillation defined from pollen and isotope records in Switzerland (Lotter et al., 1992). During the Allerød, a second short-lived cool event depicted around 13,450 cal yr B.P. in the pollen and chironomid-inferred reconstructions may be an equivalent to a cooling episode in the GRIP record (Figs. 6 and 7). A third slight cooling, centered at 12,900 cal yr B.P., can be related to the Gerzensee oscillation (Lotter et al., 1992). Chironomid and pollen-inferred reconstructions (only core 105) indicate a cooling of 2°C (Figs. 6 and 7). This result is slightly colder than the reconstructions made at Gerzensee, where pollen and cladoceran evidence suggest a cooling related to the Gerzensee oscillation of 1.2 and 0.8°C, respectively (Lotter et al., 2000).

The Preboreal oscillation (PBO), centered at 11,200 cal yr B.P., is characterized at Lake Lautrey by a cooling of 0.5° to 1°C. This agrees well with the ~0.6°C cooling derived from the Gerzensee pollen record (Birks and Ammann, 2000; Lotter et al., 2000).

Conclusions

(1) The application of different methods of pollen-based climate reconstructions to the sequences taken from Lake Lautrey have enabled us to quantify the rapid climate transitions of the Late-Glacial period of the Jura mountains.

(2) The comparison of the MAT and PFT methods shows that they provide generally similar climate signals, with the exception of the GI-1e stade. The study also shows the effectiveness of combining two different reconstruction methods, increasing the confidence in the data, and thus obtaining more precise and robust climate estimates.

(3) The inferred temperatures based on pollen and chironomids are generally in good agreement, again with the exception of the Bølling stade. In addition, there is a coherency in the timing of the changes in temperature estimates based on the two proxies suggesting, that at least at this temporal scale, there are no lags in the response of pollen and chironomids. In order to improve the pollen-based reconstructions, especially during the Bølling, it will be of interest to develop a method which uses the independent chironomid-based temperatures to constrain the pollen-based temperatures, in the same way than beetles- or organic matter-based constraints have previously been used (Guiot et al., 1993b).

(4) The comparative analysis of the results based on the three Lautrey cores have highlighted significant differences in the climate reconstructions related to the location of each core, underlining the caution that is needed when studying single cores that were not taken from deepest part of lake basins.

(5) Although a general agreement is observed between the paleotemperature estimates from various proxies (Table 3) showing an identical succession of warming (Oldest Dryas/Bølling and Younger Dryas/Preboreal transitions) and cooling (Oldest Dryas and Younger Dryas) phases, there are discrepancies depending on any particular period and/or methods used. For example, the temperature reconstructed from the Lautrey and Le Locle records for cool and warm periods are similar to those inferred from coleopteran evidence, but higher than those reconstructed at Gerzensee or in west-central Europe using other approaches. While the direction and timing of these temperature changes is increasingly well understood, the differences observed in the magnitude of the temperature changes reconstructed at Lautrey, Le Locle, Gerzensee and more generally in west-central Europe are related to differences in the methods of reconstruction and to the sensitivity or accuracy of the climatic indicators used.

Acknowledgments

This study is a part of a multi-proxy project, Eclipse (Past Environments and Climates), in which the environmental changes at Lake Lautrey will be reconstructed using pollen, lake-levels, isotopes, mineralogy, organic matter, magnetic susceptibility and chironomids records (coordinator: M. Magny). We thank the French CNRS and the Eclipse project for financial support.

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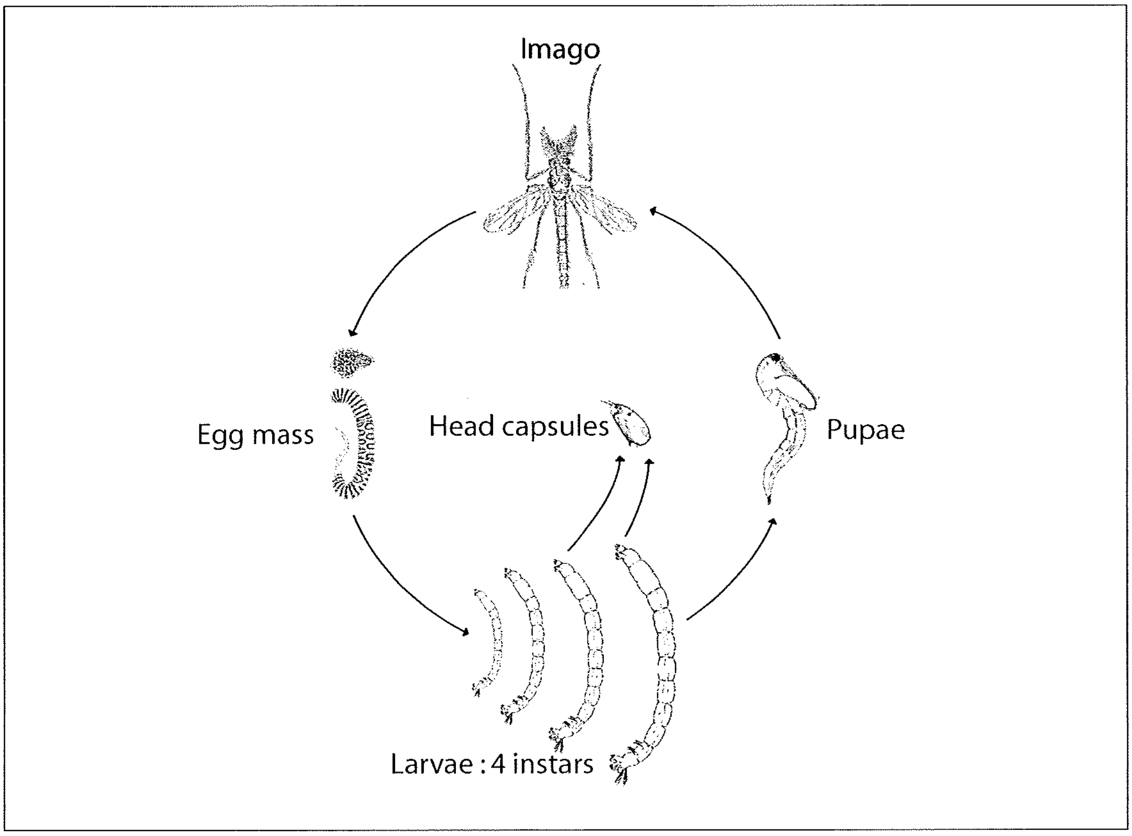
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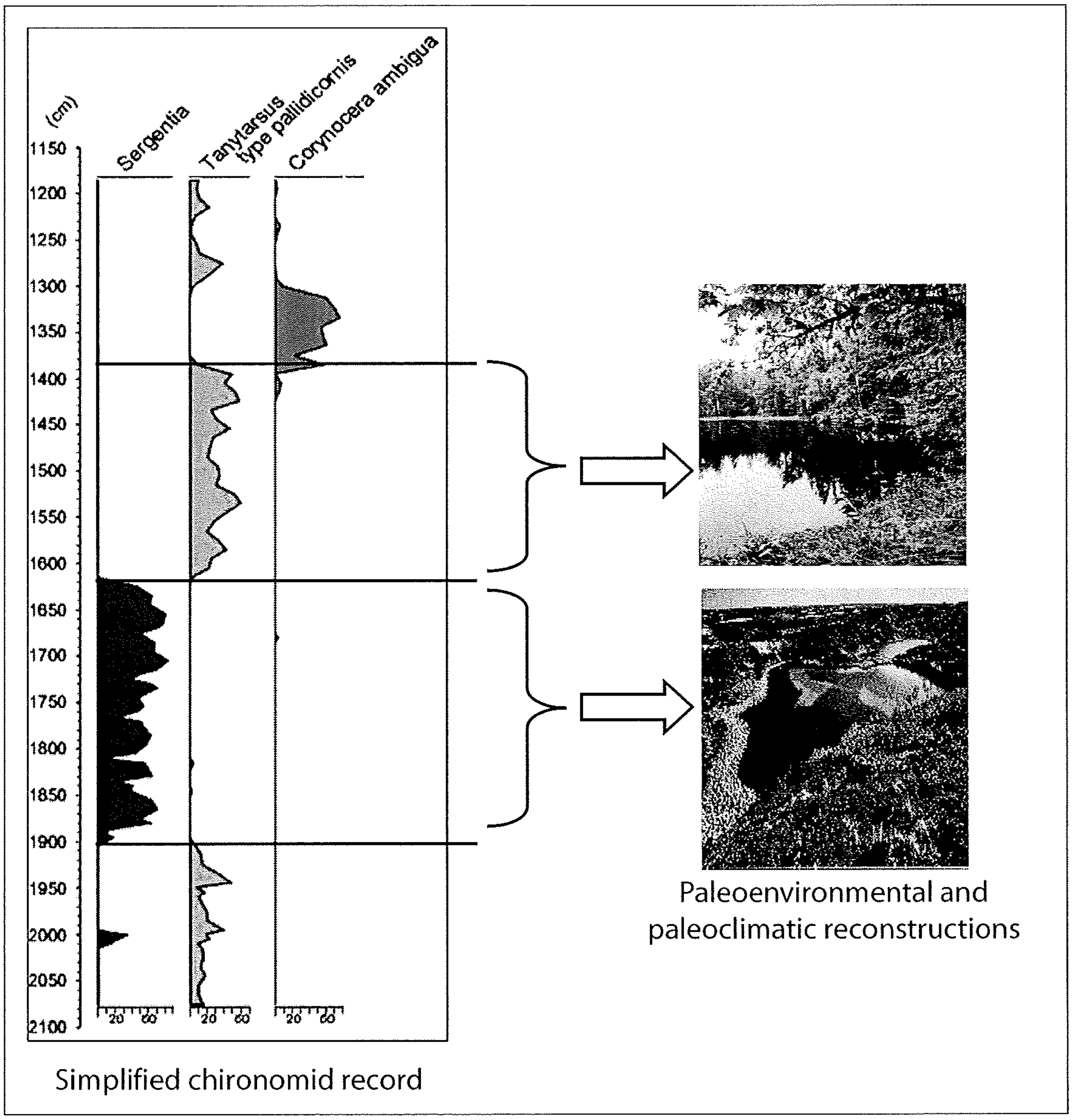
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Paleoenvironmental and paleoclimatic reconstructions
Simplified chironomid record

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Reconstruction of Late Glacial summer temperatures from chironomid assemblages in Lac Lautrey (Jura, France)

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ABSTRACT: A chironomid–July air temperature inference model based on chironomid assemblages in the surface sediments of 81 Swiss lakes was used to reconstruct Late Glacial July air temperatures at Lac Lautrey (Jura, Eastern France). The transfer-function was based on weighted averaging–partial least squares (WA-PLS) regression and featured a leave-one-out cross-validated coefficient of determination (r^2) of 0.80, a root mean square error of prediction (RMSEP) of 1.53 °C, and was applied to a chironomid record consisting of 154 samples covering the Late Glacial period back to the Oldest Dryas. The model reconstructed July air temperatures of 11–12 °C during the Oldest Dryas, increasing temperatures between 14 and 16.5 °C during the Bølling, temperatures around 16.5–17.0 °C for most of the Allerød, temperatures of 14–15 °C during the Younger Dryas and temperatures of ca. 16.5 °C during the Preboreal. The Lac Lautrey record features a two-step July air temperature increase after the Oldest Dryas, with an abrupt temperature increase of ca. 3–3.5 °C at the Oldest Dryas/Bølling transition followed by a more gradual warming between ca. 14 200 and 13 700 BP. The transfer-function reconstructs a less rapid cooling at the Allerød/Younger Dryas transition than other published records, possibly an artefact caused by the poor analogue situation during the earliest Younger Dryas, and an abrupt warming at the Younger Dryas/Holocene transition. During the Allerød, two centennial-scale 1.5–2.0 °C coolings are apparent in the record. Although chronologically not well constrained, the first of these cold events may be synchronous with the beginning of the Gerzensee Oscillation. The second is inferred just before deposition of the Laachersee tephra at Lac Lautrey and is therefore coeval with the end of the Gerzensee Oscillation. In contrast to the Greenland oxygen isotope records, the Lac Lautrey palaeotemperature reconstruction lacks a clearly defined Greenland Interstadial (GI) event 1d and the decreasing temperature trend during the Bølling/Allerød Interstadial. Copyright © 2005 John Wiley & Sons, Ltd.

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KEYWORDS: subfossil chironomids; July air temperature; France; Late Glacial; weighted averaging–partial least squares regression.

Introduction

During the Late Glacial period (ca. 15 000–11 500 BP) the global climate system shifted from the full glacial mode of the Last Glaciation to its current interglacial state. In the North Atlantic region, this period was characterised by several abrupt shifts in temperature and a number of decadal- to centennial-scale minor temperature fluctuations (e.g. Lotter *et al.*, 1992; von Grafenstein *et al.*, 1999; Johnsen *et al.*, 2001). Many qualitative and semi-quantitative temperature reconstructions covering the whole or parts of the Late Glacial period are available from Europe. For example, palaeobotanical analyses (e.g. Lotter *et al.*, 1992; Björck *et al.*, 1996), oxygen isotope ratios in lacustrine carbonates (e.g. von Grafenstein *et al.*, 1999), or lake sediment records reflecting past lake-level changes (e.g. Magny, 2001) have been used to identify periods of cooler/warmer or wetter/dryer climate. However, continuous quantitative palaeotemperature records reconstructing Late Glacial temperatures on a centigrade scale (e.g. Brooks and Birks, 2000a; Lotter *et al.*, 2000) are still rare, especially at a high temporal resolution.

Fossil remains of chironomids (non-biting midges) are increasingly being used to reconstruct quantitatively past summer temperatures. Chironomids are abundant in most

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freshwater habitats and the larval head capsules preserve well in lake sediments. Chironomid head capsules are identifiable—usually to genus or species-group level—and can thus be used to reconstruct the past chironomid fauna of lakes. In many northern temperate and subarctic regions, the distribution of lacustrine chironomid taxa is closely related to the ambient summer temperature and a number of chironomid–temperature transfer-functions based on chironomid assemblages in lake surface sediments have now been developed (e.g. Lotter *et al.*, 1997; Olander *et al.*, 1999; Brooks and Birks, 2001). These models are able to provide quantitative estimates of past summer air or water temperature based on the taxonomic composition of fossil chironomid assemblages in lake sediments. To date, Late Glacial chironomid-inferred temperature records are available from North America and Northern Europe, and cover the Younger Dryas (e.g. Walker *et al.*, 1991; Levesque *et al.*, 1993b; Brooks and Birks, 2000a; Brooks and Birks, 2000b), the Amphi-Atlantic (Gerzensee) oscillation (Levesque *et al.*, 1993a) or even the entire Late Glacial period (Brooks and Birks, 2000a). However, chironomid-based Late Glacial temperature reconstructions covering more than the end of the Younger Dryas cold period are not yet available from Central European lakes.

Here we present a chironomid-inferred July air temperature reconstruction from Lac Lautrey (Jura, France) covering most of the Late Glacial including the Oldest Dryas/Bølling transition. A chironomid–temperature inference model developed in the nearby Swiss Jura, Swiss Plateau and Northern Swiss Alps (Heiri, 2001; Heiri *et al.*, 2003b) was used to reconstruct Late Glacial July air temperatures from the fossil record.

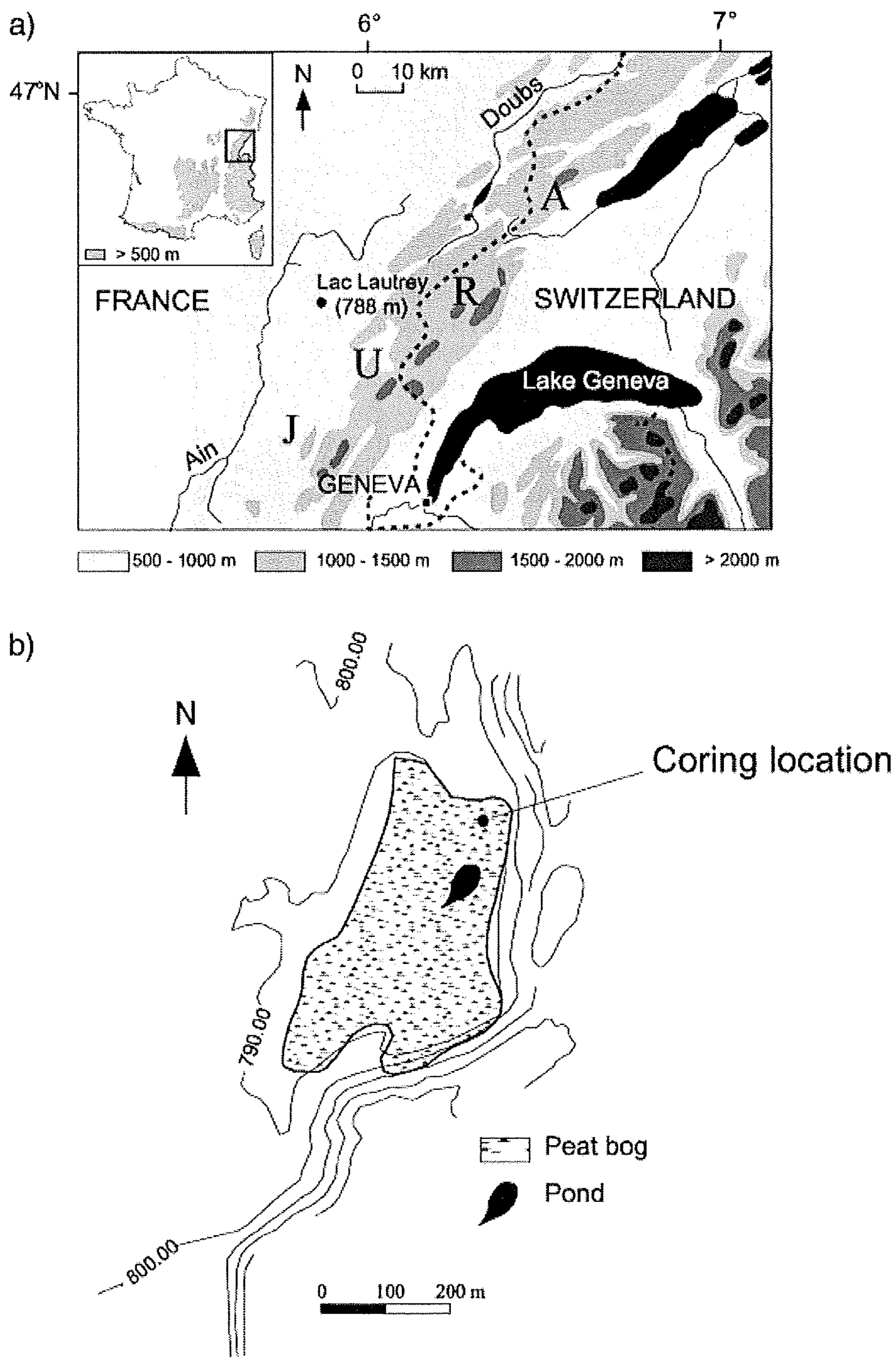

Figure 1 (a) Geographical location of Lac Lautrey in the French Jura. (b) Location of the coring site and map of the Lac Lautrey basin (following Bossuet *et al.*, 2000; Millet *et al.*, 2003)

Table 1 Biozones based on pollen analysis and lithology of the Lac Lautrey sediment record (see Magny *et al.* (submitted) for details)

Biozone	Sediment depth (cm)	Lithology	Approximate age on the Swiss Plateau/in eastern France (Magny, 2001)
Preboreal	<298	Authigenic lake marl	<11 500 BP
Younger Dryas	298–326	Biogenic lake marl (298–311 cm) and clayey silts (311–326 cm)	11 500–12 650 BP
Allerød	326–406	Silty carbonate lake marl	12 650–13 900 BP
Older Dryas	406–420	Silty carbonate lake marl	13 900–14 050 BP
Bølling	420–445	Silty (420–430 cm) and clayey (430–446 cm) carbonate lake marl	14 050–14 700 BP
Oldest Dryas	>445	Clayey silts	>14 700 BP

Study area

Lac Lautrey is a small lake situated at 788 m a.s.l. in the French Jura mountains (46° 35′ 14″ N, 5° 51′ 50″ E; Fig. 1(a)). It is located in a shallow depression in the contact area between the calcareous plateau and the folded Haut Jura mountain chain. This depression is largely filled with Quaternary lake deposits and only an open water area of ca. 1900 m^2 remains. Based on a detailed geophysical exploration of the lake deposits (Bossuet *et al.*, 2000), sediments were cored in the peat bog ca. 80 m north of the present-day lake (Fig. 1(b)) using a Russian corer (10 cm in diameter; see Magny *et al.* (2002), Millet *et al.* (2003) and Magny *et al.* (submitted) for more details on the coring and the Lac Lautrey sediment record). Pollen stratigraphic and sedimentological analyses indicate that the record contains the complete Late Glacial sequence (Table 1) and consists of lacustrine sediments deposited in a sublittoral and littoral environment (Magny *et al.*, 2002; Magny *et al.*, submitted). The present-day vegetation in the small lake catchment (2.5 km^2) consists of a dense mixed forest of coniferous (62%) and deciduous trees (38%). There are no large surface inlets and the lake is mainly fed by surface runoff from the small watershed.

Methods

From the Lac Lautrey sediment sequence 154 samples for chironomid analysis were taken at 0.5 cm to 2 cm intervals. Chironomid head capsules were extracted following the procedure described by Hofmann (1986). The head capsules were identified using Hofmann (1971) and Wiederholm (1983), and the Lac Lautrey chironomid record is described in detail by Millet *et al.* (2003). Surface sediments from 85 lakes in Northern Switzerland ranging from 420 to 2490 m a.s.l. provided the basis for a chironomid–July air temperature transfer-function. In each lake a short core was taken in the deepest part of the lake basin using a modified Kajak corer (Renberg, 1991) and chironomid head capsules were identified from the topmost 1–2 cm of the sediment core (see Heiri (2001) and Heiri *et al.* (2003b) for more details on the transfer-function). Mean July air temperature was estimated for each lake following Lotter *et al.* (1997). Owing to the lower taxonomic resolution of the Lac Lautrey chironomid record, a number of the chironomid taxa in the Swiss surface sediment data set had to be amalgamated at genus-level before regression of the chironomid–temperature transfer-function. Amalgamation was necessary for *Chironomus, Cladotanytarsus, Corynoneura, Micropsectra*, and *Paratanytarsus*. Furthermore, *Tanytarsus* I, II and III were joined to *Tanytarsus* spp. (equivalent to unidentified *Tanytarsus* and Tanytarsini in Millet *et al.*, 2003) and *Tanytarsus* IV, V and VI to *Tanytarsus pallidicornis*-type. *Tanytarsus lugens*-type and *Corynocera oliveri* were amalgamated in the Lac Lautrey record, as these taxa had not been differentiated in the Swiss data set. The transfer-function was based on weighted averaging–partial least squares regression (WA-PLS; ter Braak and Juggins, 1993) and calculated using the program CALIBRATE (S. Juggins and C. J. F. ter Braak, unpublished software) using square root transformed percentage data.

Chronology

For Lac Lautrey, 12 ^{14}C dates from wood fragments (Magny *et al.*, 2002; Millet *et al.*, 2003), δ^{18}O measurements on lacustrine carbonates and the Late Glacial/early Holocene pollen stratigraphy were available for developing an age–depth relationship (Magny *et al.*, submitted). ^{14}C dates were calibrated based on Stuiver *et al.* (1998), and, within the 95% age confidence intervals of the calibrated ^{14}C dates, shifts in the Lac Lautrey carbonate oxygen isotopes were correlated with the Greenland Ice-core Project (GRIP) δ^{18}O record at the Greenland Interstadial (GI)-1c/GI-1b, GI-1b/GI-1a, GS-1/Holocene transitions and at the Preboreal Oscillation (at 11 200 GRIP yr BP) (Björck *et al.*, 1998; Johnsen *et al.*, 2001). This correlation was furthermore constrained by the Laachersee tephra (at 345 cm; dated to ca. 12 880 BP by Litt *et al.*, 2001) and by the Oldest Dryas, Bølling, Older Dryas, Allerød, Younger Dryas, and Preboreal pollen zones (Begeot, 2000; Magny, 2001; Magny *et al.*, 2002; Magny *et al.*, submitted; Table 1). All these pollen zones are clearly delimited in the Lac Lautrey sediments (Magny *et al.*, submitted) and therefore allow a correlation with other Late Glacial pollen records from the region and with associated lacustrine oxygen isotope records (e.g. Lotter *et al.*, 1992). For the Lac Lautrey age–depth relationship the age estimates for Late Glacial pollen zone boundaries provided by Magny *et al.* (2001) were used (Table 1). These age estimates are in agreement with the correlation between the GRIP oxygen isotope record and lacustrine oxygen isotope records from the Swiss Plateau proposed by Björck *et al.* (1998). Sediment sections between the correlated time markers were assigned ages based on linear interpolation. Age estimates of the Oldest Dryas silt-clay deposits are based on extrapolation of the sedimentation rates of the early Bølling carbonate lake marl and should therefore be interpreted with caution. All further references to ages for the Lac Lautrey record are based on this age-scale (which is described in detail in Magny *et al.* (submitted)), and are given in GRIP years BP (AD 1950 = 0 BP), except where otherwise indicated.

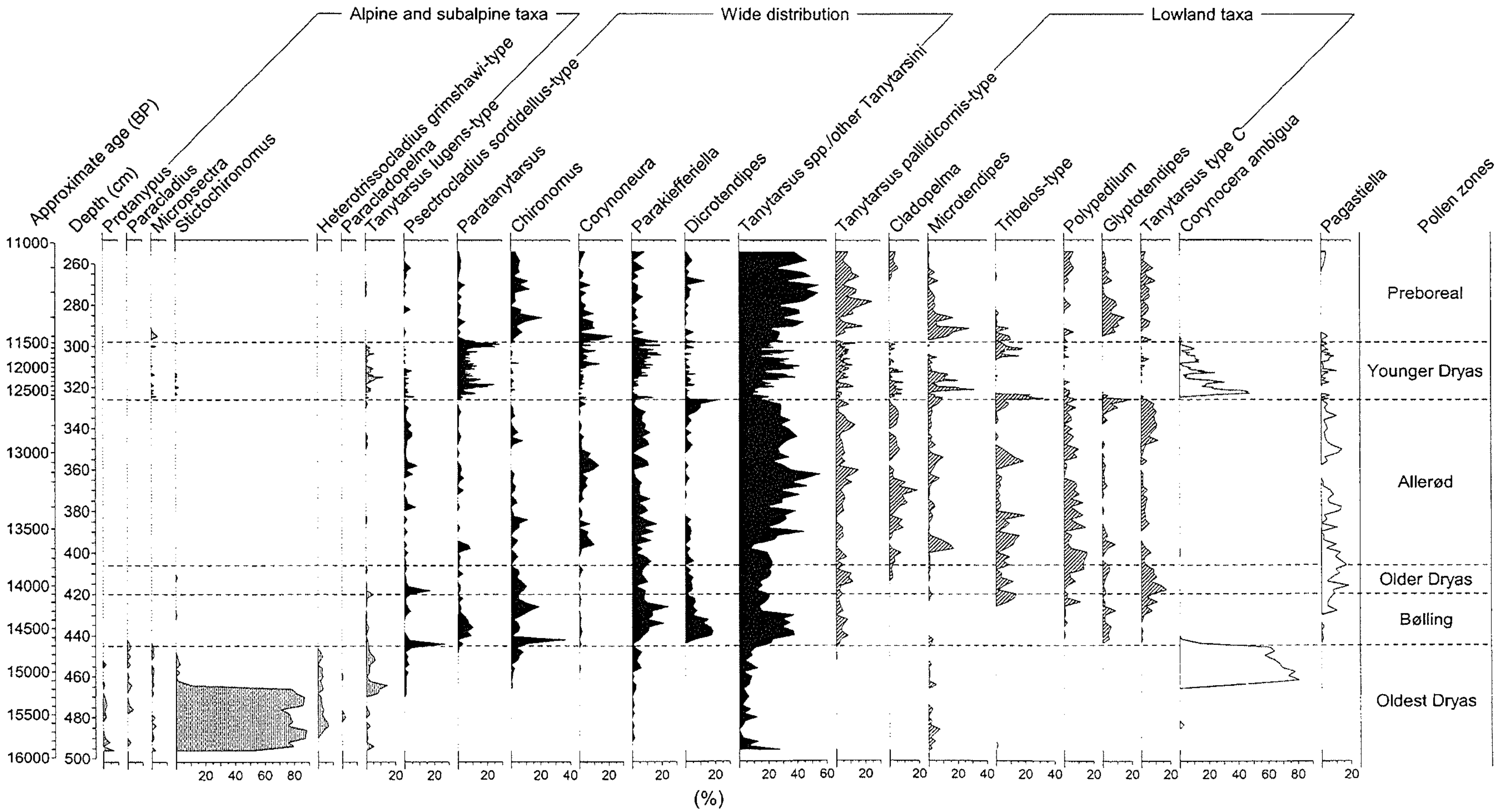

Figure 2 The Lac Lautrey chironomid record. The chironomids are classified into alpine/subalpine taxa, lowland taxa and taxa with a wide distribution in respect to temperature based on their occurrence in small lakes in Northern Switzerland (*Corynocera ambigua* and *Pagastiella* are not represented in this data set; Fig. 3). With the exception of the alpine/subalpine chironomids, only chironomid taxa with at least 15% abundance in at least one sample of the record are shown. The pollen zonation follows Magny *et al.* (submitted). (See Table 1)

Results

The Lac Lautrey chironomid record

Chironomid assemblages in the lowest part of the Lac Lautrey sediment record are dominated by *Stictochironomus*, with *Heterotrissocladius grimshawi*-type, *Protanypus*, *Paracladius*, *Tanytarsus lugens*-type, *Tanytarsus* spp. and *Microtendipes* present at lower abundances (Fig. 2). At 464 cm sediment depth, there is an abrupt shift to chironomid assemblages dominated by *Corynocera ambigua*. At 447.5 cm depth, *C. ambigua* disappears from the record and a number of other chironomid taxa increase in abundance, including *Chironomus*, *Dicrotendipes*, *Tanytarsus pallidicornis*-type, *Parakiefferiella*, *Glyptotendipes*, *Tanytarsus* type C, *Tanytarsus* spp. and *Pagastiella*. The next distinct change in the chironomid record is apparent at 324 cm depth, where *C. ambigua* reappears and chironomid taxa such as *Tanytarsus lugens*-type, *Microtendipes* and *Paratanytarsus* show higher abundances. At the same boundary other chironomids such as *Tanytarsus* spp., *Tribelos*-type, *Glyptotendipes* and *Tanytarsus* type C decline in abundance. Between 324 and 298 cm there is a gradual shift in the abundances of some taxa, including a decrease in *C. ambigua* and *Microtendipes* and an increase in *Parakiefferiella*, *Tanytarsus* spp. and *Tribelos*-type. At 298 cm, a number of taxa disappear from the Lac Lautrey sediments, including *C. ambigua*, *Pagastiella*, *Cladopelma* and *Tanytarsus lugens*-type, and others such as *Parakiefferiella* and *Paratanytarsus* decrease in abundance.

Of the dominant chironomids in the Late Glacial record of Lac Lautrey, *C. ambigua* and *Pagastiella* have not been found in the Northern Swiss lake sediment samples available for the development of the chironomid–temperature transfer-function. *Stictochironomus*, *Paracladius*, *Protanypus*, *Micropsectra*, *Tanytarsus lugens*-type and *Paracladopelma* are largely restricted to high altitude, cold Swiss lakes (Fig. 3). *Psectrocladius sordidellus*-type is most abundant in cold lakes. However, the taxon is found in lower abundances in lakes over the whole temperature gradient. A number of taxa, including *Paratanytarsus*, *Chironomus*, *Corynoneura*, and *Parakiefferiella*, show a wide distribution in respect to summer temperature. *Dicrotendipes*, and *Tanytarsus* spp. occur over the whole temperature range. However, they are not very common at the cold end of the temperature gradient. The remaining taxa, including *Cladopelma*, *Microtendipes*, *Tanytarsus pallidicornis*-type, *Polypedilum* and *Glyptotendipes*, are largely restricted to habitats with July air temperatures higher than 12 °C, although they are occasionally found in cooler lakes (Fig. 3). *Tribelos*-type and *Tanytarsus* type C are typically found in Swiss lowland lakes at air temperatures above 14 °C.

Transfer-function development

Of the 85 surface sediment samples available for the regression of the chironomid–July air temperature inference model (Heiri, 2001), four were excluded from the calculations since they originate from lakes used as reservoirs or lakes susceptible to flooding from a nearby river, or are dominated by a taxon otherwise rare in Swiss lake sediments. The remaining 81 samples were used to calculate a chironomid–July air temperature transfer-function based on WA-PLS (ter Braak and Juggins, 1993; ter Braak *et al.*, 1993). The WA-PLS model with two components featured a leave-one-out cross-validated root mean square error of prediction (RMSEP) of 1.53 °C and a coefficient of determination (r^2) of 0.80 (Fig. 4). The inclusion of a third component led to only slightly improved error statistics (RMSEP 1.51 °C; r^2 0.81) and the simpler two-component model was therefore retained. Although a number of chironomid taxa in the Swiss surface sediment data set had to be amalgamated to be consistent with the taxonomic resolution of the Lac Lautrey record, the model's performance is only slightly reduced compared with the original chironomid–temperature inference model with a higher taxonomic resolution (RMSEP 1.51 °C, r^2 0.81; Heiri *et al.*, 2003b).

Chironomid-inferred temperatures

Below the Oldest Dryas/Bølling transition at 445 cm sediment depth, July air temperatures fluctuating around 11–12 °C are inferred in the Lac Lautrey record (Fig. 5(a)). Inferred temperatures rise abruptly at 445 cm depth and reach a plateau of 14.0–14.5 °C in the earliest Bølling. At ca. 425 cm depth, temperatures increase again to reach values fluctuating around 16.5–17.0 °C during most of the Allerød. Two centennial-scale temperature decreases of ca. 1.5–2.0 °C are inferred at ca. 360 and 345 cm sediment depth. At the beginning of the Younger Dryas (326 cm sediment depth), reconstructed July air temperatures decrease from values fluctuating around 17.0–17.5 °C to ca. 15 °C. The inferred July air temperatures continue to decrease more gradually between 326 and 315 cm to reach minimum values of about 14 °C at ca. 315 cm sediment depth. During the rest of the Younger Dryas, temperatures gradually increase to ca. 15 °C at 298 cm sediment depth. At the Younger Dryas/Holocene transition, July air temperatures rise abruptly by about 1.5 °C and fluctuate around 16.5 °C in the earliest part of the Holocene. For a single chironomid sample at 270 cm sediment depth, a July air temperature value of 14 °C is inferred.

Discussion

Reliability of the inferred temperatures

Most of the Late Glacial chironomid assemblages analysed in the Lac Lautrey record are dominated by taxa at present abundant in the Swiss Alps, the Swiss Plateau and the Jura mountains (Figs 2, 3). During two phases, however, a significant proportion of the chironomid head capsules in the sediments belonged to taxa not present in the extant chironomid assemblages used to calibrate the chironomid–July air temperature transfer-function (Fig. 5(b)). This non-analogue situation was mainly due to two chironomid taxa, *Corynocera ambigua* and, to a lesser extent, *Pagastiella* (Fig. 2). *Corynocera ambigua* has traditionally been considered a cold stenothermous chironomid species, since its remains are often found in sediments corresponding to Late Glacial cold intervals (Brodersen and Lindegaard, 1999). However, in arctic and subarctic environments the distribution of *C. ambigua* does not show a clear relationship with temperature (Olander *et al.*, 1999), or the species is most abundant at the warm end of the temperature gradient (Porinchu and Cwynar, 2000). Recently, high abundances of *C. ambigua* have been described from warm, shallow and eutrophic Danish lakes (Brodersen and Lindegaard, 1999), casting further doubt on the cold stenothermous character of the species. In the 85 surface sediment samples available for calibration of the chironomid–temperature

Figure 3 Distribution of the chironomid taxa dominating the Lac Lautrey record in the surface sediment samples used to develop the chironomid–temperature transfer-function. The samples are ordered according to the mean July air temperature at the lakes studied (see Lotter *et al.* (1997) and Heiri (2001) for lake codes and for further information on the sites)

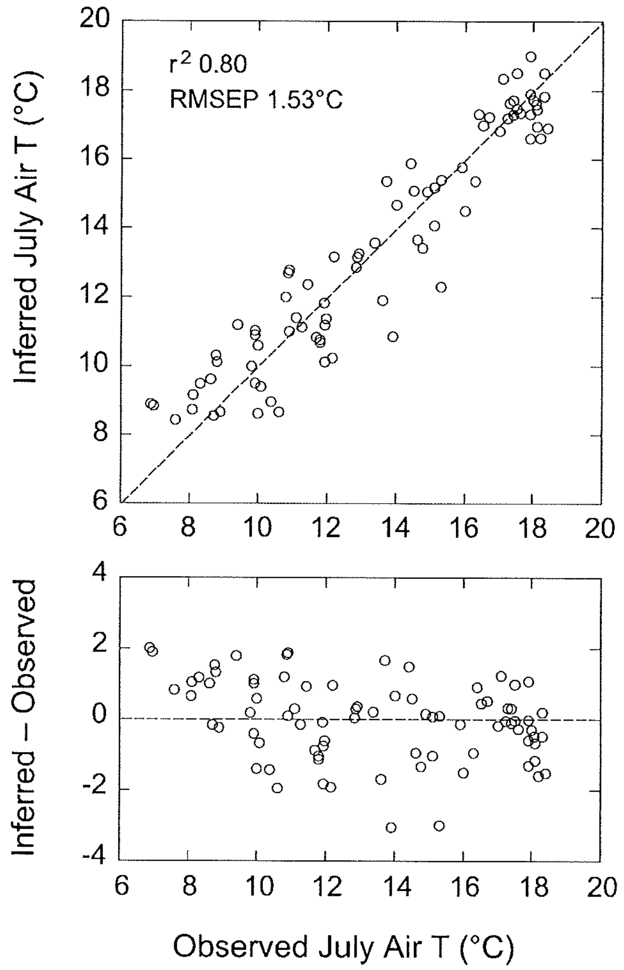

Figure 4 Inferred temperatures and prediction residuals of the chironomid–July air temperature transfer-function plotted versus the observed temperatures. RMSEP indicates the leave-one-out cross-validated root mean square error of prediction and r^2 the cross-validated coefficient of determination

transfer-function, not a single specimen of *C. ambigua* has been found, indicating that the species is not part of the extant chironomid fauna of the Northern Alpine region. *Pagastiella* occurs over a wide temperature range and tends to be more abundant in warmer lakes (e.g. Olander *et al.*, 1999; Larocque *et al.*, 2001; Porinchu *et al.*, 2002). Like *C. ambigua*, *Pagastiella* is absent from the surface sediment samples used to develop the chironomid–temperature transfer-function. However, live specimens have been collected in Switzerland (Lods-Crozet, 1998) and head capsules have been found in surficial lake sediments in the Central Alpine region (O. Heiri, unpublished data).

Chironomid taxa in the Lac Lautrey sediments not represented in the chironomid–temperature transfer-function do not contribute to chironomid-inferred temperatures. As a consequence, temperature inferences from samples with a high proportion of *C. ambigua* and *Pagastiella* are based on only a part of the chironomid fauna and on a comparatively low number of chironomid specimens (Fig. 5(b)). Low count sizes can add to the variability of and add a bias to chironomid-inferred temperatures (Heiri and Lotter, 2001). For the Lac Lautrey chironomid record, this implies that inferred temperatures at the end of the Oldest Dryas (ca. 15 100–14 700 BP) and at the beginning of the Younger Dryas (ca. 12 650–12 200 BP) should be interpreted with caution, as they are based on samples with a high proportion of *C. ambigua* and *Pagastiella* (Fig. 5).

The surface sediment samples used to calibrate the chironomid–July air temperature transfer-function were obtained from the deepest parts of the sampled lake basins (Heiri, 2001). The Lac Lautrey sediment sequence, on the other hand, represents sublittoral/littoral deposits. A detailed study of subfossil chironomid assemblages in Norwegian lakes indicates that a bias in inferred temperatures is possible if chironomid–temperature inference models calibrated on deep-water sediments are applied to nearshore assemblages (Heiri *et al.*, 2003a). However, in shallow lakes such as Lac Lautrey, this bias has been estimated to be ca. 0.4–0.5 °C and is therefore relatively small compared with the Late Glacial temperature changes inferred in the record.

Late Glacial temperature development

The Lac Lautrey record is unique in providing a high-resolution chironomid-based temperature record which covers most of the Late Glacial period, including the Oldest Dryas. As expected, the record follows previous Late Glacial temperature reconstructions in Europe in inferring the strongest temperature changes at the Oldest Dryas/Bølling (ca. 14 700 BP), Allerød/Younger Dryas (ca. 12 650 BP) and Younger Dryas/Holocene transitions (ca. 11 500 BP). During the Oldest Dryas (before 14 700 BP), July air temperatures around 11–12 °C are inferred at Lac Lautrey. Chironomid assemblages between 464 and 445 cm sediment depth are dominated by a high proportion of chironomids not included in the chironomid–temperature transfer-function and, therefore, chironomid-inferred temperatures are based on a comparatively low number of counts per sample. Nevertheless, chironomid-inferred temperatures remain relatively stable during the Oldest Dryas interval (Fig. 5). Based on a large number of Late Glacial pollen records, Renssen and Isarin (2001) reconstruct minimum July air temperatures around 14–15 °C (at sea level elevation) for Central Europe during the Oldest Dryas. Assuming an altitudinal lapse rate of 6 °C km^{-1} for July air temperature (Livingstone *et al.*, 1999), this value suggests minimum July air temperatures of 9.5–10.5 °C at the altitude of Lac Lautrey (788 m a.s.l.), and agrees well with the chironomid-inferred temperatures (Fig. 5(a)). At the Oldest Dryas/Bølling transition, chironomid-inferred temperatures rise abruptly from ca. 11 °C to reach values around 14–14.5 °C at ca. 14 600 BP. This chironomid-inferred temperature increase of 3–3.5 °C is in good agreement with the pollen-inferred minimum July temperature increase of ca. 3–4 °C reconstructed for Central Europe by Renssen and Isarin (2001) for this climate transition. However, based on hindcasts of Late Glacial climate by an atmospheric general circulation model (GCM), Renssen and Isarin (2001) conclude that their pollen-inferred July air temperature reconstruction most probably underestimates Bølling summer temperatures in Central Europe. The Lac Lautrey temperature record indicates a two-step increase of July air temperatures after the Oldest Dryas/Bølling transition, with the abrupt temperature increase followed by a more gradual rise in July air temperatures during the second half of the Bølling. This is in contrast to a number of other European reconstructions of Late Glacial climate, which suggest an abrupt, single-step increase to the warm temperatures of the Bølling/Allerød Interstadial (e.g. Eicher and Siegenthaler, 1976; Lotter *et al.*, 1992; Brooks and Birks, 2000a). In the Lac Lautrey record, the first of these two temperature increases (at ca. 14 700 BP) is characterised by the disappearance of most alpine and subalpine chironomid taxa (Figs 2, 3). At the same time, a number of chironomid taxa typical of extant lowland lakes colonise Lac Lautrey

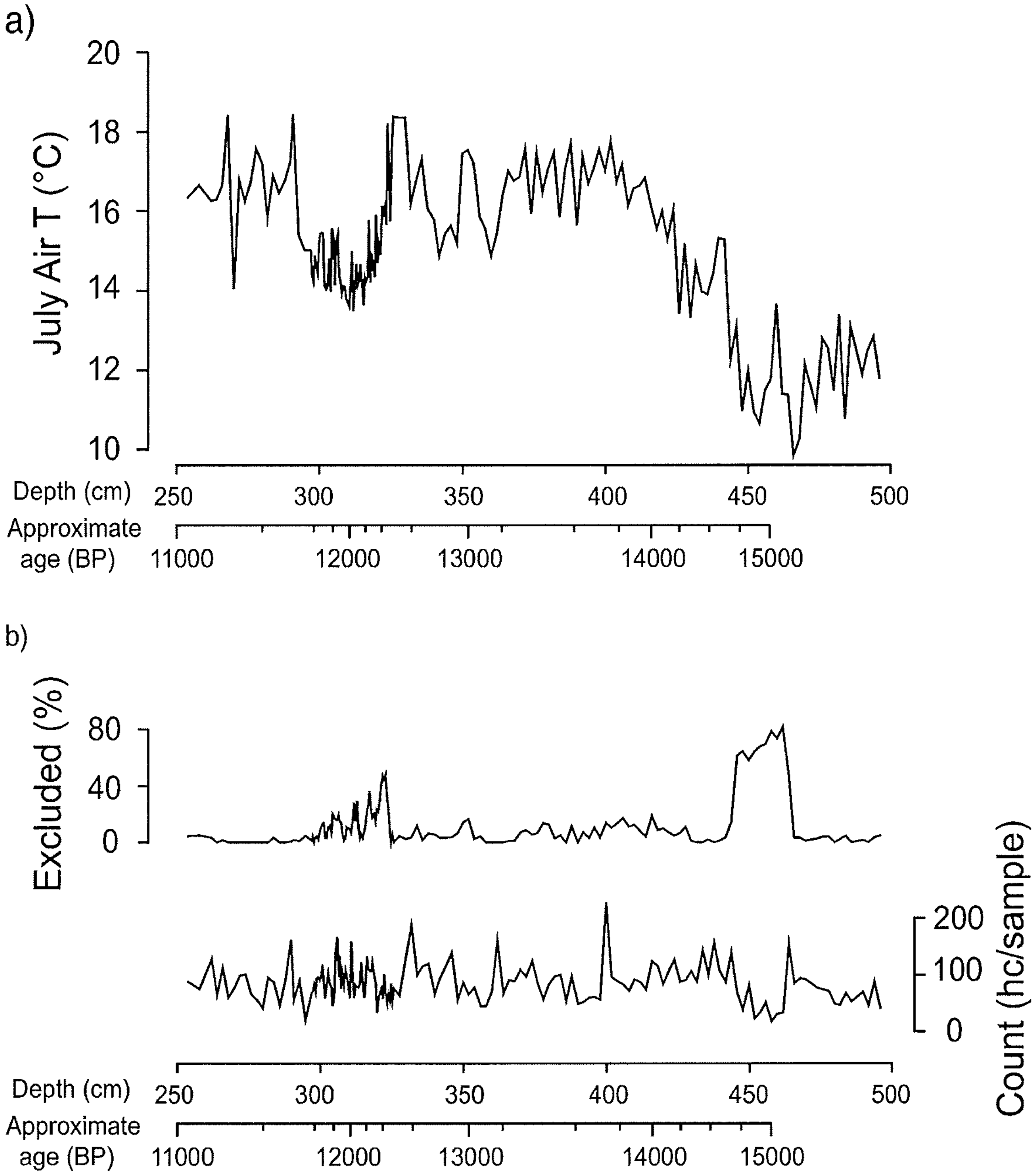

Figure 5 (a) Chironomid-inferred July air temperatures from Lac Lautrey. (b) Percentage of identified chironomids in the record not represented in the transfer-function and count size of the fossil samples excluding these chironomid taxa

(e.g. *Tanytarsus pallidicornis*-type and *Glyptotendipes*). During the second more gradual temperature increase between ca. 14 200 and 13 700 BP, a number of these taxa increase in abundance, e.g. *Polypedilum*, and *Tanytarsus* type C, and further lowland taxa colonise the lake, including *Tribelos*-type and *Cladopelma*. The two distinct waves of lowland chironomid taxa colonising Lac Lautrey raise the question of whether the cooler chironomid-inferred temperatures at the beginning of the Bølling are an artefact due to immigration lags of the warm-adapted chironomid species. However, chironomids are able to actively fly and seek out suitable habitats for oviposition, and can therefore rapidly colonise lakes. Furthermore, Lac Lautrey is a mountain lake and it is expected that most temperate chironomid taxa were already present in the adjacent lowlands during the Oldest Dryas. A number of the lowland chironomids present in Lac Lautrey in the earliest Bølling increase in abundance as *Cladopelma* and *Tribelos*-type colonise the lake (Fig. 2), and this suggests that the shift in chironomid assemblages between ca. 14 200 and 13 700 BP was due to changing environments rather than to immigration lags.

Chironomid-inferred July air temperatures fluctuate around 16.5–17.0 °C during most of the Allerød (ca. 13 900–12 650 BP) with the highest temperatures of 17.5–18.0 °C reconstructed just before the beginning of the Younger Dryas. Temperature records based on fossil cladoceran and pollen assemblages from Northern Switzerland reconstruct summer temperatures 3.5–5.5 °C cooler than present-day values for the end of the Allerød (Lotter *et al.*, 2000). In contrast, the chironomid-inferred July air temperatures at Lac Lautrey (788 m a.s.l.) during the Allerød are distinctly higher than the present-day values of, for example, 15 °C mean July air temperature measured in 1997 at 800 m a.s.l. in Northern Switzerland (Livingstone *et al.*, 1999).

Two centennial-scale temperature decreases of 1.5–2.0 °C centred on ca. 13 200 and 12 900 BP are apparent in the Lac Lautrey palaeotemperature record (Fig. 5(a)). During the

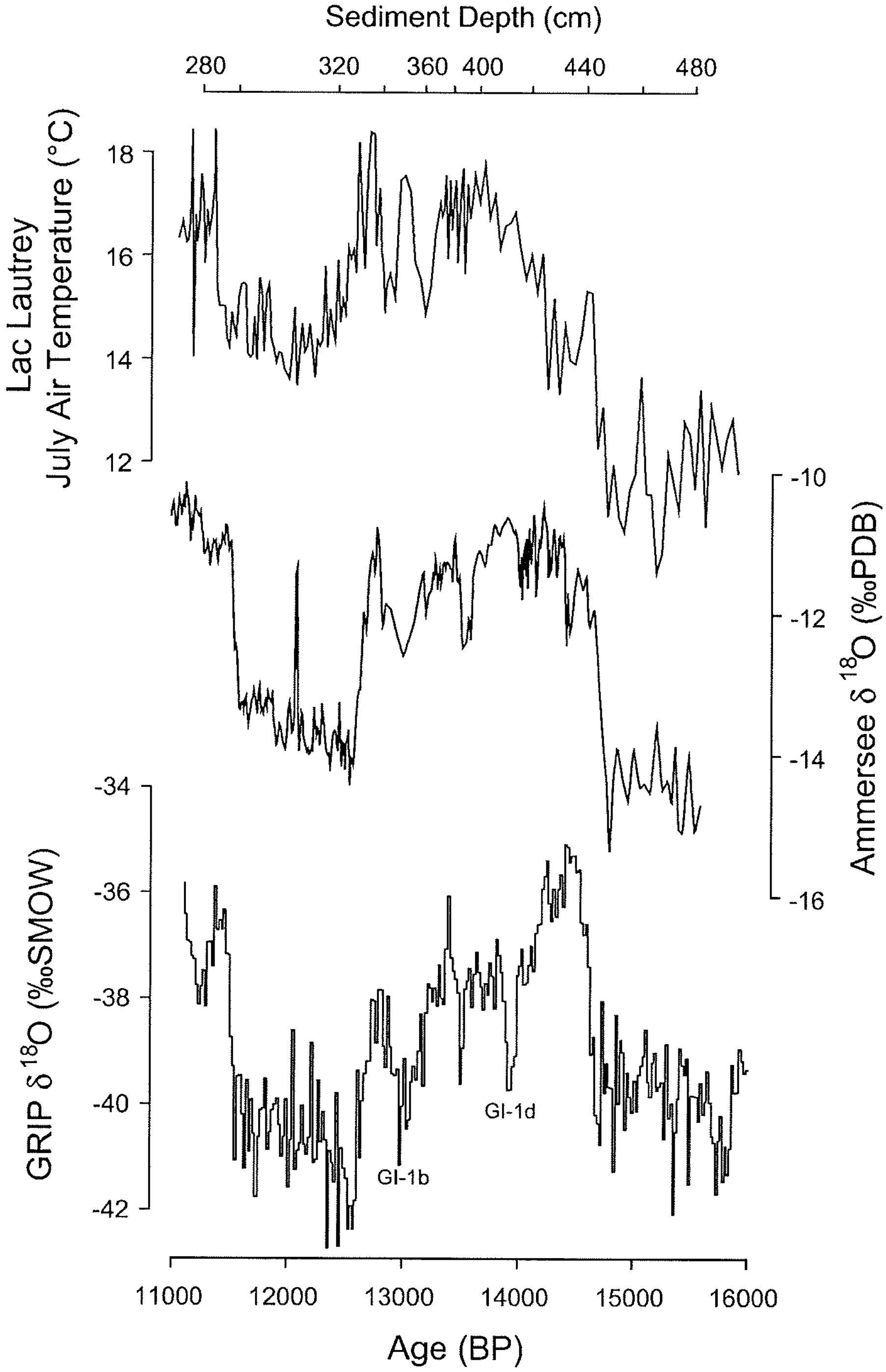

Figure 6 Chironomid-inferred July air temperatures from Lac Lautrey, oxygen isotope ratios measured on ostracod remains from Ammersee, southern Germany (von Grafenstein *et al.*, 1999), and oxygen isotope ratios in the GRIP ice core (Johnsen *et al.*, 2001). The Greenland Interstadial (GI) events 1b and 1d are indicated following Björck *et al.* (1998)

Bølling/Allerød Interstadial (ca. 14 700–12 650 BP), three centennial-scale temperature fluctuations have been reported from Central Greenland, based on the Greenland ice-core oxygen isotope records (Johnsen *et al.*, 1992; Dansgaard *et al.*, 1993; Björck *et al.*, 1998; Johnsen *et al.*, 2001; Fig. 6). Evidence has been presented indicating that the first of these events, centred on ca. 14 000 BP (Greenland Interstadial event 1d ; GI-1d), has affected European climate as well (Björck *et al.*, 1998) and is equivalent to the Aegelsee Oscillation described from Northern Switzerland based on palynological and oxygen isotope analyses of lake sediments (Lotter *et al.*, 1992). A second, minor cold oscillation in the Greenland oxygen isotope records is apparent at ca. 13 500 BP. Upwelling records from the Cariaco Basin, offshore Venezuela (Hughen *et al.*, 1998), and oxygen isotopes in a lake sediment record from Southern Germany (von Grafenstein *et al.*, 1999) suggest that this event affected the whole North Atlantic climate system. The third temperature decrease (GI-1b) is centred on ca. 13 000 BP. Further palaeoclimatic evidence for this cooling event, the Gerzensee Oscillation or Amphi-Atlantic Oscillation, has been presented from both sides of the North Atlantic (Lotter *et al.*, 1992; Levesque *et al.*, 1993a; Brooks and Birks, 2000a; Bedford *et al.*, 2004). The younger of the centennial-scale coolings in the Lac Lautrey temperature record begins shortly before the deposition of the Laachersee tephra at 345 cm sediment depth. Since this tephra layer has been dated to 12 880 BP (Litt *et al.*, 2001), this suggests that the second Bølling/Allerød cooling in the Lac Lautrey temperature record is synchronous with the end of the Gerzensee Oscillation (and GI-1b). The earlier of the two centennial-scale coolings in the Lac Lautrey record is inferred significantly later than the Bølling/Allerød transition, which is apparent at 406–420 cm sediment depth in the pollen record (Magny *et al.*, 2002; Magny *et al.*, submitted). In Central Europe, the age of the Bølling/Allerød transition has been estimated to ca. 13 900–14 050 BP (Magny, 2001) and this, therefore, indicates that the older centennial-scale cooling recorded at Lac Lautrey cannot be equivalent to GI-1d. Based on the age–depth relationship of the Lac Lautrey sediment record described in Magny *et al.* (submitted), the age of the older chironomid-inferred cooling period in the Allerød has been estimated to ca. 13 200 BP. Oxygen isotope ratios measured on carbonates in the Lac Lautrey sediments indicate a decrease in $\delta^{18}O$ values synchronous with this cooling (Magny *et al.*, submitted). However, since the Lautrey oxygen isotope record contains a considerable amount of noise during the Allerød it is difficult to assess whether this $\delta^{18}O$ reduction is equivalent to the beginning of the Gerzensee Oscillation. The amplitude of the two chironomid-inferred Allerød coolings is close to the prediction error of the chironomid–July air temperature inference model of 1.53 °C. Given the additional uncertainties of applying a transfer-function based on mid-lake samples to a littoral/sublittoral sediment record, these two inferred July air temperature fluctuations should perhaps best be interpreted with caution.

The Allerød/Younger Dryas boundary in the Lac Lautrey record is marked by a decrease in chironomid-inferred July air temperature from values fluctuating around 17.0–17.5 °C at ca. 12 700 BP to temperatures around 14 °C at ca. 12 300 BP. This temperature decrease is not as abrupt as inferred in many other Late Glacial palaeotemperature records from Europe (e.g. Eicher and Siegenthaler, 1976; von Grafenstein *et al.*, 1999; Brooks and Birks, 2000a, 2000b), although cladoceran- and pollen-based summer air temperature reconstructions in the Swiss lowlands also suggest a more gradual temperature decrease (Lotter *et al.*, 2000). Chironomid-assemblages in the Lac Lautrey sediments between ca. 12 600 and 12 200 BP contain a large proportion of chironomids not included in the chironomid–temperature inference model (Figs 2, 5). Therefore, this gradual decrease in inferred July air temperatures at the Allerød/Younger Dryas boundary and during the first part of the Younger Dryas should be interpreted with caution, especially since a very abrupt change in chironomid assemblages is apparent in the sediment record at ca. 12 650 BP (Fig. 2). During the second part of the Younger Dryas, July air temperatures of ca. 14–15 °C are inferred at Lac Lautrey. This agrees well with GCM-based late Younger Dryas July temperature estimates of 20 °C at sea level for Central Europe (15.3 °C if corrected to the altitude of Lac Lautrey; Renssen and Isarin, 2001), and chironomid-inferred July air temperatures of ca. 10.5 °C reconstructed at 1515 m a.s.l. in the Northern Swiss Alps (14.9 °C if corrected to the altitude of Lac Lautrey; Heiri *et al.*, 2003b). However, based mainly on palynological data, Renssen and Isarin (2001) reconstructed a minimum July air temperature of 16 °C at sea level in Central Europe (11.3 °C at the altitude of Lac Lautrey). Using cladoceran remains and pollen in the sediments of Gerzensee (603 m a.s.l., on the Swiss Plateau), Lotter *et al.* (2000) inferred mean summer (June, July, August) temperatures around 10 °C for the end of the Younger Dryas. At present, July air temperatures in Northern Switzerland are highly correlated to and ca. 1 °C higher than mean summer temperatures. The Younger Dryas summer air temperatures inferred at Gerzensee and corrected to the altitude of Lac Lautrey would therefore be equivalent to ca. 10 °C July air temperature and significantly cooler than the chironomid-inferred temperatures from Lac Lautrey. With the exception of the highly uncertain early Younger Dryas temperatures at ca. 12 600–12 200 BP, our palaeotemperature record shows the gradual increase of temperature during the Younger Dryas that is apparent in a number of European temperature reconstructions (e.g. von Grafenstein *et al.*, 1999; Brooks and Birks, 2000a).

At Lac Lautrey, inferred July air temperatures increase by ca. 1.5 °C at the Younger Dryas/Holocene transition. This agrees well with the summer temperature increase of ca. 2 °C inferred by cladocerans and pollen in Northern Switzerland (Lotter *et al.*, 2000), the July air temperature increase of 1.5 °C inferred by chironomids in the Northern Swiss Alps (Heiri *et al.*, 2003b), and the increase of 2 °C of the temperature of the warmest month inferred by pollen records and lake-level fluctuations from Le Locle in the Swiss Jura mountains (Magny *et al.*, 2001). However, using pollen data and GCM simulations, Renssen and Isarin (2001) infer a July air temperature increase of ca. 3–4 °C for the Younger Dryas/Holocene transition in Central Europe.

Based on the observation that temperature records on both sides of the North Atlantic show a similar sequence of Late Glacial climate events, it has been suggested that the high-resolution Greenland ice-core oxygen isotope records could provide a template against which to examine Late Glacial climate change in Europe (e.g. Björck *et al.*, 1998; von Grafenstein *et al.*, 1999; Brooks and Birks, 2000a). Major shifts in chironomid-inferred July air temperatures from Lac Lautrey show a good agreement with the GRIP oxygen isotopes (Johnsen *et al.*, 1992, 2001) (Fig. 6). However, a number of differences are apparent in the records. Before 14 700 BP, inferred July air temperatures at Lac Lautrey are significantly cooler than during the Younger Dryas. Furthermore, the temperature increase after 14 700 BP takes place in two discrete steps in the chironomid-inferred temperatures, which constrasts with the single-step, abrupt temperature increase inferred by the Greenland oxygen isotope record. This discrepancy may be partially explained by seasonal differences in the Late Glacial temperature development, as GCMs suggest significant differences between the development of summer and winter temperatures in the Late Glacial (Renssen and Isarin, 2001). However, a two-step increase in Bølling temperatures has also been reconstructed based on a high-resolution oxygen isotope record measured on ostracod valves from Ammersee, southern Germany (von Grafenstein *et al.*, 1999) (Fig. 6). The Ammersee record features a number of further agreements with the chironomid-inferred temperatures from Lac Lautrey, including cooler inferred temperatures during the Oldest Dryas than during the Younger Dryas, the absence of a clearly resolved centennial-scale cooling event at ca. 14 000 BP (GI-1d), two distinct centennial-scale coolings during the Allerød, and the absence of the decreasing temperature trend during the Bølling/Allerød Interstadial inferred by the Greenland oxygen isotope records.

Summary and conclusions

(1) We used a chironomid–temperature transfer-function, based on chironomid assemblages in 85 lakes in Northern Switzerland, to reconstruct July air temperatures from a Late Glacial chironomid record from Lac Lautrey (French Jura mountains).
(2) Chironomid assemblages indicate July air temperatures of ca. 11–12 °C during the Oldest Dryas, increasing temperatures between ca. 14 and 16.5 °C during the Bølling, temperatures of ca. 16.5–17.0 °C during most of the

Allerød, temperatures of ca. 14–15 °C during the Younger Dryas and temperatures of ca. 16.5 ° during the Preboreal.

(3) In contrast to a number of other European Late Glacial temperature reconstructions, the Lac Lautrey chironomid record suggests a two-step July air temperature increase at the Oldest Dryas/Bølling transition, with temperatures of 14.0–14.5 °C in the earliest Bølling (ca. 14 600–14 200 BP). Furthermore, the cooling at the Allerød/Younger Dryas transition takes place more gradually than in most other European Late Glacial temperature records. However, chironomid-inferred temperatures during the early Younger Dryas (ca. 12 650–12 200 BP) were strongly affected by non-analogue problems and this gradual temperature decrease should therefore be interpreted with caution.

(4) Two centennial-scale coolings during the Allerød are apparent in the Lac Lautrey chironomid-inferred temperatures. The age of the younger of these is well constrained by the presence of the Laachersee tephra, indicating that the cooling is synchronous with the end of the Gerzensee Oscillation described from the Swiss Plateau (Lotter *et al.*, 1992). The older cold oscillation is chronologically not as well constrained but may be related to the beginning of the Gerzensee Oscillation.

(5) Oldest Dryas and Bølling temperatures inferred at Lac Lautrey are within the range of pollen-based reconstructions of Late Glacial July air temperatures in Central Europe (Renssen and Isarin, 2001). In contrast to cladoceran- and pollen-based palaeotemperature records from the adjacient Swiss Plateau (Lotter *et al.*, 2000), the Lac Lautrey chironomid reconstruction suggests July air temperatures above modern-day values at the end of the Allerød. Chironomid-inferred July air temperatures during the Younger Dryas agree well with chironomid-based estimates from the Northern Swiss Alps (Heiri *et al.*, 2003b) and GCM-based reconstructions (Renssen and Isarin, 2001), but are warmer than cladoceran- and pollen-inferred July air temperature estimates for Central Europe (Lotter *et al.*, 2000; Renssen and Isarin, 2001). The Younger Dryas/Holocene increase of July air temperatures in the Lac Lautrey chironomid record of 1.5 °C is of similar magnitude to the increase inferred by a number of other proxy records in the region (1.5–2.0 °C; Heiri *et al.*, 2003b; Lotter *et al.*, 2000; Magny *et al.*, 2001), but in conflict with the pollen-inferred temperature estimates and GCM results of Renssen and Isarin (2001). Additional Late Glacial summer temperature reconstructions will be necessary to resolve the differences between the available reconstructions.

(6) A number of differences with the Greenland oxygen isotope records are shared by the Lac Lautrey chironomid-inferred temperature record and the high-resolution oxygen isotope record from Ammersee, Southern Germany (von Grafenstein *et al.*, 1999). Both the Lac Lautrey and Ammersee records feature a two-step increase in early Bølling temperatures, lack the gradual temperature decrease during the Bølling/Allerød Interstadial apparent in the Greenland oxygen isotope records, do not feature a clearly resolved Greenland Interstadial event 1d, and infer cooler temperatures during the Oldest Dryas than during the Younger Dryas (Fig. 6). Thus, the Lac Lautrey record supports the differences in millennial-scale climate trends between Europe and Greenland proposed by von Grafenstein *et al.* (1999) for the Greenland Interstadial 1.

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Pollen-inferred palaeoclimate reconstructions in mountain areas: problems and perspectives

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ABSTRACT: Transfer functions are an efficient tool for the quantitative reconstruction of past climate from low to mid-elevation pollen sites. However, the application of existing methods to high-altitude pollen assemblages frequently leads to unrealistic results. In the aim of understanding the causes of these biases, the standard 'best modern analogue' method has been applied to two high-altitude pollen sequences to provide quantitative climate estimates for the Lateglacial and Holocene periods. Both pollen sequences (Laghi dell'Orgials, 2130 m, SW aspect and Lago delle Fate, 2240 m, E aspect) are located in the subalpine belt, on opposing sides of the St. Anna di Vinadio Valley (Italian Maritime Alps). Different results were obtained from the two sequences. The largest differences occurred in palaeotemperature reconstruction, with notable differences in both the values and trends at each site. These biases may be attributed to: (1) a lack of high elevation 'best modern analogues' in the database of modern samples; (2) the problem of pollen taxa that have multiple climatic significance; (3) problems related to the complexity of mountainous ecosystems, such as the phenomenon of uphill transport of tree pollen by wind. Possible improvements to the reconstruction process are discussed. Copyright © 2006 John Wiley & Sons, Ltd.

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KEYWORDS: Italian Alps; modern analogue technique; mountain areas; palaeoclimate reconstruction; palynology.

Introduction

Pollen data represent the most widely available quantitative record of past climates (COHMAP Members, 1988) and are an efficient tool for the reconstruction of vegetation and its responses to abrupt climate changes (e.g. Birks and Ammann, 2000; Peteet, 2000; Wick, 2000; Tinner and Lotter, 2001). Palynology is the most common tool for Quaternary palaeoecology and palaeoclimate reconstructions (Huntley and Prentice, 1993; Tarasov *et al.*, 1999; Allen *et al.*, 2000; Lotter *et al.*, 2000; Peyron *et al.*, 2000; Magny *et al.*, 2001; Seppä and Birks, 2001, 2002; Davis *et al.*, 2003) and for palaeoenvironmental reconstruction for archaeology (Richard, 1995, 1997; Gauthier 2002), as it permits reconstruction of both vegetation and climate and provides evidence of agriculture and human landscape modification (Behre, 1981, 1988; Miras *et al.*, 2004; Court-Picon *et al.*, 2005).

However, whilst the reconstruction of past climate yields good results when applied to pollen sequences from low to mid-elevation sites, the application of current methods to high-altitude pollen sequences often gives unreliable results (Huntley and Prentice, 1988; Cheddadi *et al.*, 1997; Klotz *et al.*, in press). This may be due to several possible causes, for example the lack of modern pollen samples from high elevation sites in the modern pollen databases used for palaeoclimate reconstruction, wind-driven uphill transport of tree pollen into the subalpine and alpine zones, or as a result of the complex relationships between vegetation belts and relative pollen percentages in spectra from elevated sites, which depend on local physiographic conditions (slopes, exposure to dominant winds, aspect) (Brugiapaglia *et al.*, 1998; Müller *et al.*, 2000; Ortu, 2002). This complicates the problem of interpretation of high-elevation pollen records, as it is not possible to express relationships between variations in pollen percentages and elevation by a linear function.

These problems limit our understanding of the reaction of montane vegetation to past and therefore future climate change, an understanding that is essential for several reasons:

1 Montane vegetation has been shown to be particularly sensitive to climate changes (Birks and Ammann, 2000; Wick, 2000). The strong altitudinal climatic gradient that characterises mountain areas results in a steep ecological gradient, so several ecotones occur in a small area. This results in an amplification of global climate signal (Beniston *et al.*, 1997) so vegetation response to climatic changes is more pronounced at higher altitudes than in the lowland (Birks and Ammann, 2000).

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2. Mountainous areas are prized as sources of high biodiversity. Recent climate variations have had a particularly strong effect on montane ecosystems (Houghton *et al.*, 2001). Being able to assess the impacts of climatic fluctuations on the past environment biodiversity in these mountain regions will help in choosing suitable management strategies in the near future.
3. In the lowlands, sediments in lakes and peat bogs are often disturbed by human activities, whereas anthropogenic disturbance is limited to deforestation and grazing at high elevations, and lakes and peat bogs are generally undisturbed (Catalan *et al.*, 2002; Livingstone *et al.*, 2005). Montane pollen sequences therefore represent ideal archives for the study of past climate changes.

The difficulties in reconstructing vegetation history from palynological records in mountain areas have long been known (Barthélémy and Jolly, 1989; David, 1993b). The use of complementary techniques is often suggested to validate pollen-based palaeoclimate reconstructions (David, 1997; Birks, 2003). In recent studies, pollen analysis is often accompanied with a macrofossil study, in order to aid in the interpretation of pollen percentages. However, whilst macrofossil analysis can prove the local presence of taxa, it cannot prove their absence (Birks, 2003) and it is of little use for quantitative reconstructions. This technique requires a large sample size in order to obtain quantitative estimates of different taxa percentages in the vegetation and this information is in any case very local: hygrophilous vegetation surrounding lakes and peat bogs often prevents the macrofossils of the dominant vegetation cover from reaching the lake (Birks, 2003). Further, the large number of existing palynological studies has given rise to regional and continental databases containing fossil pollen data from the alpine chain (European Pollen Data Base, Alpine Pollen Data Base), which may be used for comparative purposes or regional scale studies. In contrast, macrofossil databases are still not widely available as these studies remain rare in comparison to pollen ones.

Since one of the aims of this paper is an understanding of biases in pollen-based palaeoclimate reconstruction techniques which have been used in large-scale studies, no complementary techniques are considered here. In order to examine the reliability of quantitative estimates of the Lateglacial and Holocene climate from pollen sequences located at high elevation in the Alps, we have compared palaeoclimatic reconstructions from two pollen sequences from high-elevation sites on opposing sides of a valley in the Italian south-western Alps.

Pollen analysis of the two sequences showed a common regional record of vegetation development during the Lateglacial and the Holocene, with differences in pollen percentages of some taxa due to local parameters (Ortu *et al.*, 2005). Based on radiocarbon dates, similar pollen records and existing studies, fossil samples recording the same vegetation phases in the different cores have been correlated by qualitative analysis (for more details see extensive discussion in Ortu *et al.*, 2005). Correspondence analysis (Benzecri, 1973) was used to test whether, despite local differences, the correlation obtained with a quantitative method is in agreement with those obtained qualitatively. The standard 'modern analogue technique' (MAT) (Guiot, 1990) was subsequently applied to the two pollen sequences to estimate palaeotemperature and palaeoprecipitation during the Lateglacial and the Holocene (e.g. Guiot *et al.*, 1989; Cheddadi *et al.*, 1998). The comparison of results helped identify the problems linked to pollen-based palaeoclimatic reconstruction in mountainous areas, with the aim of improving the reconstruction methodology.

Study area

The two sites (Fig. 1) are situated on opposing sides of the St. Anna di Vinadio Valley, on the Italian side of the southwestern Alps (Ortu *et al.*, 2005). Laghi dell'Orgials (2240 m a.s.l.) is a small peat bog on the western side of the valley, situated above the timberline formed by larch and surrounded by pastures and shrubs. Lago delle Fate (2130 m a.s.l.) is situated on the eastern side of the valley, and surrounded by open larch formations, pastures and shrubs. A full description of the two sites is given in Ortu *et al.* (2005). The vegetation cover and climate are presently very similar at both sites: the mean temperature of the coldest month (T_c) is about −7 °C, the mean temperature of the warmest month (T_w) is about 10 °C, the mean annual temperature (T_{ann}) about 1.4 °C and total annual precipitation (P_{ann}) about 1150 mm yr^{-1} (Cagnazzi and Marchisio, 1998).

Methods

Pollen analysis

Pollen analysis was carried on a sediment core taken using a Russian corer (Jowsey, 1966) at each site. Nine ^{14}C AMS dates were obtained on vegetal macro-remains by the Radiocarbon Dating Centre of n' Claude Bernard Lyon 1' University and the Poznan Radiocarbon Laboratory. The dates are given as calendar years before present (BP) (Table 1). The calibration of dates follows Reimer *et al.* (2004).

The two pollen diagrams, which are presented in summary form (Fig. 2), show comparable regional vegetation phases. The correlation of the same climate and vegetation phases is based on the radiocarbon dates and by correlation of the pollen zones showing similar vegetation dynamics. A full description of diagrams is given in Ortu *et al.* (2005). Local taxa typically developing in alpine peat bogs (Cyperaceae, Juncaceae) and lakes (*Sparganium/Typha*; *Pediastrum* algae), which are known for giving irregular pollen percentages that cannot be directly compared to vegetation cover or used for palaeovegetation reconstruction (Berglund and Ralska-Jasiewiczowa, 1986) have been excluded from the pollen total sum for the reconstruction of palaeovegetation surrounding each site (local vegetation: 0–20 m around the site; Heim, 1970). These percentages often obscure those of other taxa which are dominant in the vegetation surrounding the studied site, and whose development is primarily determined by climate. These taxa have been excluded from the pollen total sum for the palaeovegetation reconstruction (Ortu *et al.*, 2005) and they were not used for the climate reconstruction, as their development is linked to processes concerning the peat bog and lake evolution, such as filling or eutrophication.

Correspondence analysis

Correspondence analysis (Benzecri, 1973) was applied to both present and fossil pollen data from the two sites in order to group together quantitatively similar pollen spectra. The program selects a certain number of factorial axes in order of decreasing importance. The importance of an axis is based on the amount of the total variance it explains. We have retained the first three axes selected by the analysis (explaining 78.52% of the total variance) for the construction of ordination biplots, in order to show the grouping of similar samples (Fig. 3).

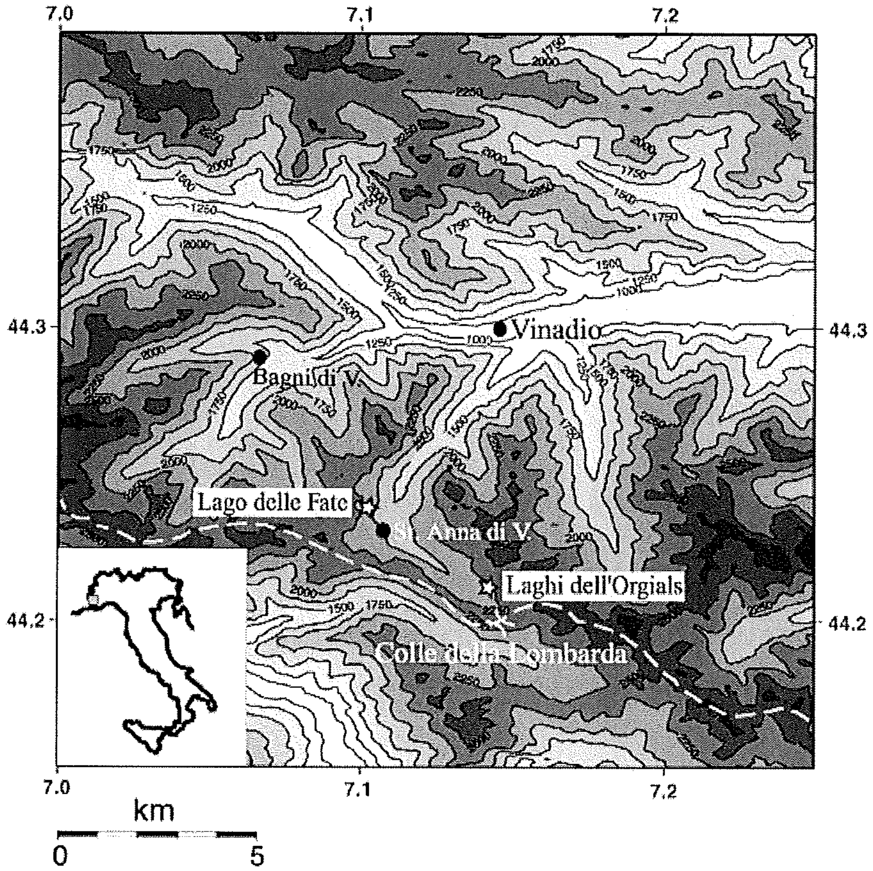

Figure 1 Location of the studied sites in the St. Anna di Vinadio Valley: Lago delle Fate (2130 m, E aspect) and Laghi dell'Orgials (2240 m, SW aspect)

Pollen-based palaeoclimate reconstruction

The standard 'modern analogue technique' (MAT) (Guiot, 1990) was applied to the two pollen sequences to estimate palaeotemperature and palaeoprecipitation during the Lateglacial and the Holocene (e.g. Guiot *et al.*, 1989; Cheddadi *et al.*, 1998). In the MAT method, eight similar modern pollen spectra are selected from a modern sample database for each fossil pollen assemblage and their climate is averaged to provide an estimate of the climate corresponding to the fossil assemblage (Guiot, 1990). Each modern sample site was assigned a modern climate based on interpolation from station data using an artificial neural network (Guiot *et al.*, 1996). The search for analogues is based on the squared chord distance (Overpeck *et al.*, 1985), using an equation to find a set of s (here $s = 8$) modern analogues of the fossil spectra (chord distance < 0.4). The quality of the reconstruction is expressed by the climate homogeneity of the s analogues, as well as the mean chord distance (Table 2). The reconstructed climate value for each fossil spectrum is the distance-weighted mean of the climate values associated with the s best analogues. Instead of a unique standard deviation, the error bars are calculated using the average of the analogues above and below the weighted average value. Whilst it is acknowledged that this may underestimate the errors, we have retained this method: (a) to provide consistency with existing studies; (b) to provide palaeoclimate error estimates that are a better reflection of the distribution of the chosen modern analogues. In the present work, two different modern databases have been used. The first database used included 1328 modern pollen spectra (Peyron *et al.*, 1998). This was then subsequently improved by the addition of new pollen spectra from Spain, Tibetan high-altitude steppe vegetation and Scandinavian pioneer vegetation. Samples in which the climate-vegetation may have been biased owing to

Table 1 Results of ^{14}C dating: codes, material, identification, depth, age of the dated samples

Sample	Dated material	Identification	Depth (cm)	Age (^{14}C yr BP)	Calibrated age (2σ) (BP)	$\delta\ ^{13}C$ (‰)
Lago delle fate						
LYON-1597(GRA-19336)	Wood	*Alnus* sp.	133	6480 ± 100	7572-7245	−28,03
LYON-1596(GRA-19335)	Wood	*Salix* sp.	141	8850 ± 60	10174-9730	28,45
LYON-1595(GRA-19334)	Wood	*Salix* sp.	171	9300 ± 60	10609-10283	−27,92
LYON-1594(GRA-19331)	Wood	*Salix* sp.	214	9660 ± 60	11206-10779	—
Laghi dell'Orgials						
LYON-1599(OXA)	Vegetal remains	—	59	1875 ± 35	1885–1719	−28,26
Poz-7107	Gyttja	—	142	4670 ± 35	5473–5315	—
Poz-7108	Gyttja	—	234	7820 ± 40	8723–8515	—
LYON-1611(OXA)	Bryophytae	*Drepanocladus*	359	9783 ± 68	11351–11077	—
LYON-1598(GRA-19337)	Bryophytae	*fluitans*	406	20930 ± 130	25552–24719	—

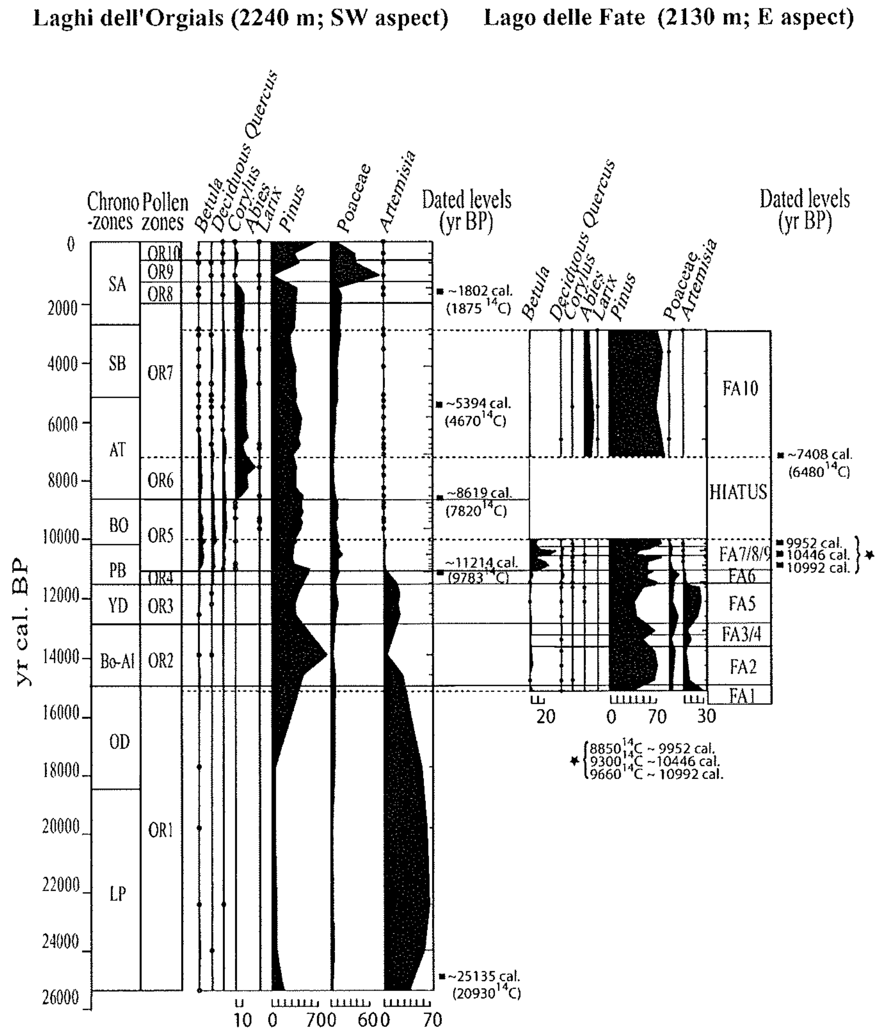

Figure 2 Synthetic pollen diagrams at the two sites: pollen zones and chronological correlation. Chronozones: LP, Late Pleniglacial; OD, Oldest Dryas; Bo, Bölling; Al, Alleröd; YD, Younger Dryas; PB, Preboreal; BO, Boreal; AT, Atlantic; SB, Subboreal; SA, Subatlantic. The diagrams are displayed on a calibrated timescale. Ages of the dated levels are expressed in ^{14}C and cal. yr BP

anthropogenic landscape transformation were deleted, giving a database of 868 modern pollen spectra (Peyron *et al.*, 2005). Surface pollen samples were converted into biomes (Prentice *et al.*, 1996), in order to carry out a comparison with the biomes simulated by the BIOME1 model (Prentice *et al.*, 1992). Pollen samples inconsistent with the potential vegetation at the sites of their collection were removed.

To evaluate the reliability of the method, the climate parameters for each surface sample were estimated using the other modern samples. The difference between present-day climate data at the pollen sites and the estimated climate at each site indicates the quality of the method (Table 2). As expected, the application of MAT to the improved database has higher coefficients of correlation between the observed and the estimated parameters.

Table 2 Correlation coefficients between observed and reconstructed values of climate parameters obtained of MAT approach to both modern pollen databases. Climate parameters: T_c: mean temperature of the coldest month; T_w: mean temperature of the warmest month; T_{ann}: mean annual temperature; P_{ann}: mean annual precipitation

Climatic parameter	Database 1 Correlation coefficient	Database 1 RMSE	Database 2 Correlation coefficient	Database 2 RMSE
T_c (°C)	0.955	3.9	0.97	3.6
T_w (°C)	0.856	2.5	0.922	2.3
T_{ann} (°C)	0.947	2.6	0.972	2.2
P_{ann} (mm/yr)	0.797	178.7	0.88	149.1

The MAT was used to reconstruct the mean annual temperature (T_{ann}), total annual precipitation (P_{ann}), the mean temperature of the coldest month (T_c) and the mean temperature of the warmest month (T_w). Results from the two sequences are displayed on a chronological scale (Figs 4–6) obtained by the interpolation of relative and calibrated radiocarbon dates by the program GPalWin (Goeury, 1988). We were unable to obtain ^{14}C dates for this period, owing to the mineralogical nature of sediment (with the exception of the bottom of the OR1 zone). There is, however, a large body of existing studies available for the western Alps, in which the transitions between the major Lateglacial fluctuations (Oldest Dryas—Bölling–Older Dryas—Alleröd–Younger Dryas) are dated. The regional radiocarbon-dated pollen stratigraphy has been used to date the Lateglacial sequences. This allows a direct comparison to be made between the two curves, in spite of differences in sedimentation rate and the presence of hiatus.

Pollen-based palaeoclimate reconstruction constrained by biomes

Tests using the standard MAT method showed an excessive heterogeneity between the *s* best analogues due to the effects of human disturbance and the lack of perfect modern analogues. It is possible to reduce this bias by selecting modern analogues using biome scores. These scores are calculated as a total of all taxa representing a chosen biome (Prentice *et al.*, 1996). It is assumed that if a taxon is missing in the modern samples, it will be replaced by another from the same biome, thus maintaining a similar biome score despite taxonomic differences. Using this method each modern and fossil pollen assemblage has been attributed to a biome. Eleven biomes are described for Europe: tundra, cold deciduous forest, taiga, cool conifer forest, cold mixed forest, temperate deciduous forest, cool mixed forest, warm mixed forest, xerophytic wood/shrub, steppes. During the selection of the analogues, the biome assigned to the fossil assemblage is compared to the modern samples and modern analogues for which the biomes are not consistent are rejected (Guiot *et al.*, 1993).

Results

Qualitative pollen zones correlation

A full description of the two pollen sequences and of the vegetation history interpreted from them is given in Ortu *et al.* (2005). The correlation of pollen zones is shown in Fig 2. In the description of results and their discussion, Latin names of taxa are used for pollen percentages, while English names are used for their interpretation as vegetation cover. The two sequences show similar pollen spectra recording the same vegetation phases at a regional scale. The bottom of the Orgials sequence records the end of the Pleniglacial. Lateglacial fluctuations are recorded at both sites: the Oldest Dryas (pollen zones OR1 and FA1) and the Younger Dryas (OR3 and FA5) are characterised by high steppe taxa (dominated by *Artemisia*) percentages. The Older Dryas fluctuation (FA3), separating the Bölling (FA2) and the Alleröd (FA4) periods, may be observed in the Fate sequence. The Bölling–Alleröd is recorded in a single pollen zone at Orgials (OR2) and characterised at both sites by high *Pinus* percentages (80%). High *Pinus* percentages also characterise pollen zones at the beginning of the Holocene (OR4, FA6). The importance of this taxon in pollen spectra is due to its morphological adaptations to wind transport, which make it dominant in pollen percentages when no other taxon producing important pollen rates is locally present. The over-representation of *Pinus* in pollen spectra is well known, and the absence of pinewoods surrounding high-altitude sites during the Bölling–Alleröd and the beginning of the Holocene has been well described (David, 1993b). At Fate, the development of a birch wood surrounding the site between 11 000 and 9950 cal. yr BP is recorded in pollen zones FA7, FA8, FA9, and equally observed at Orgials (OR5). Some evidence is seen of the development of a lowland oak forest during the same period at both sites (FA7, 8, 9; OR5), with higher pollen percentages at Orgials, where there was no surrounding forest to filter transported pollen. A pine wood developed at Fate after 10 450 cal. yr BP, gradually replacing birch in the subalpine belt (FA9, FA10). The development of fir in the mountainous and subalpine belts is dated around 8600 cal. yr BP at Orgials (OR6), while it is not recorded at Fate because of a hiatus between 9950 and 7400 cal. yr BP. The FA10 zone at Fate indicates the presence of an open pine wood at this site after 7400 cal. yr BP with larch and fir. The development of these taxa in lower belts is recorded at Orgials during this period by uphill transport of pollen by the wind (Ortu *et al.*, 2005). The Orgials site appears to have been situated above the timberline throughout the Lateglacial and the Holocene, while the Fate site had been surrounded by a birch wood since 11 000 cal. yr BP and by a pine wood, following the disappearance of birch (Ortu *et al.*, 2005). The common background that characterises the two sequences represents regional scale vegetation dynamics. This allows the correlation of climate-induced vegetation cover variations at the two sites, from the Lateglacial to present.

Correspondence analysis on pollen data

Figure 3 shows the results of the correspondence analysis of the pollen data from the two sites. The biplot based on the two first axes (Fig. 3(a) and (b)), shows that in some instances, the pollen zones belonging to the same vegetation phases are not grouped together. The first axis is formed on the negative side by steppic taxa and on the positive side by arboreal taxa, and represents a moisture gradient from dry (negative) to wet (positive).

The second axis represents variations in values of *Betula* (negative) and *Abies* (positive). Pollen zones OR7, 8, 10 and FA7, 8, 9 are split into separate groups on the second axis despite belonging to the same chronological period. This split is driven by the locally high values of *Abies* at Orgials and locally high values of *Betula* at Fate. All pollen spectra that have high abundances of non-local taxa are gathered near the axes' origin, as they are not well-enough characterised by the pollen taxa associated with the first two axes (FA6; OR4, 5) or they show a mixture of taxa coming from different vegetation belts (FA10: *Abies* + *Pinus* dominant; OR6: *Abies* + oak mixed forest taxa). The biplot based on the first and the third axes (Fig. 3(c) and (d)), in which *Betula* has a less important role, shows a good correlation between the pollen zones corresponding to the same vegetation phases in the two sequences (OR4–5/FA6–8; OR6–7/FA9–10). The Oldest (OR1) and Younger Dryas (OR3; FA5) pollen spectra form two distinct groups on the negative side of the first axis. The spectra belonging to the FA1 and FA3 zones, which correspond to the end of the Oldest Dryas and to the Older Dryas, are grouped with the FA2 group, owing to low steppe taxa (*Ephedra*, Chenopodiaceae) percentages, and the presence of several alpine pasture taxa (Labiatae, *Centaurea*, *Armeria*, Boraginaceae). The tree-rich Holocene pollen samples are gathered on the positive side of the first axis. The first phases of the Holocene (11 500–9950 cal. yr BP), corresponding to the development of pine and birch at high elevation and of the oak deciduous forest in the lowland (pollen zones OR4–5 and FA6–8), are grouped together on the positive side of the third axis. The following Holocene phases, corresponding to the development of the fir wood (pollen zones OR6–7 and FA10) are grouped on the negative side of the third axis. Modern spectra (taken from the area around the sites) are grouped with the fossil ones characterised by high percentages of non-local arboreal pollen, near the axes' origin.

The results of the multivariate analysis of the two sequences show that the pollen samples belonging to the same vegetation (and climatic) phases may be grouped using a qualitative interpretation and the chronology, despite local differences.

Palaeoclimatic reconstruction

The results of the MAT method (Guiot, 1990) show notable differences between the two sites (Fig. 4). At both sites, during the Lateglacial, T_c values are lower than present. However, at

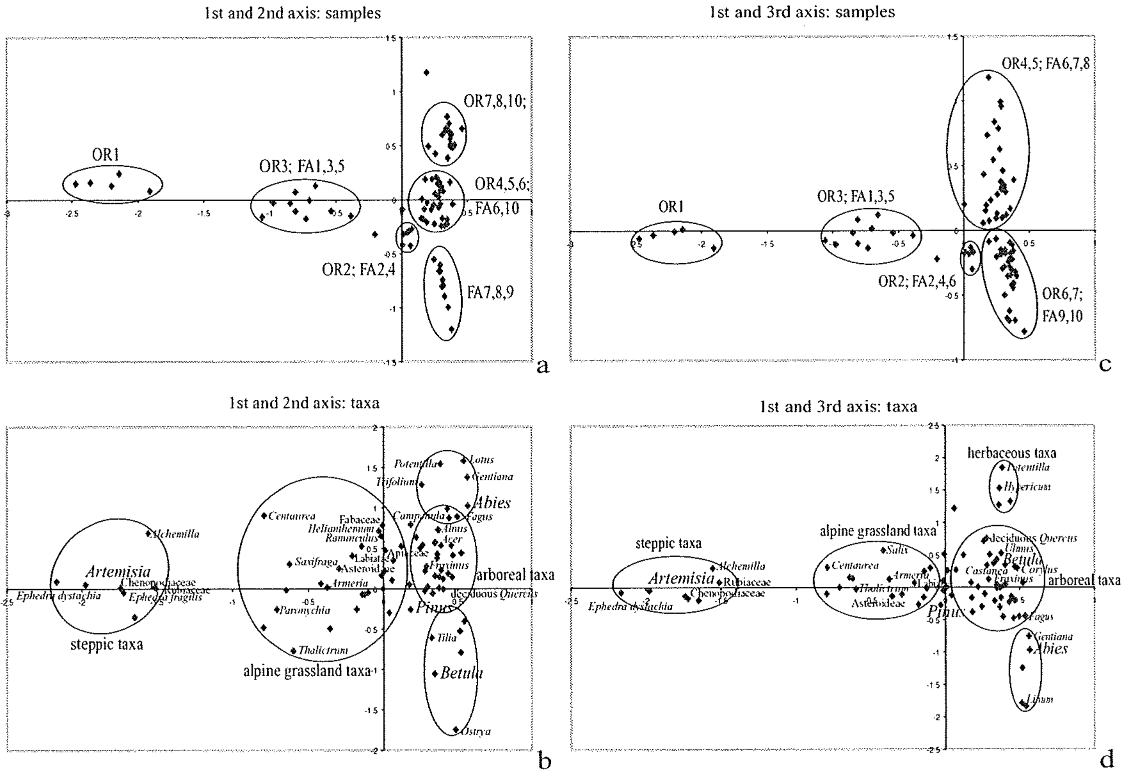

Figure 3 Results of correspondence analysis on fossil pollen samples at the two sites. (a) Dispersion of pollen samples: groups for the percentages of 77 taxa in 86 samples. Horizontal axis = 1st factorial axis (43.42% of the total variance), vertical axis = 2nd factorial axis (20.32% of the total variance). (b) Dispersion of pollen taxa: groups for the percentages of 77 taxa in 86 samples. Horizontal axis = 1st factorial axis (42.44% of the total variance), vertical axis = 2nd factorial axis (33.59% of the total variance). (c) Dispersion of pollen samples: groups for the percentages of 77 taxa in 86 samples. Horizontal axis = 1st factorial axis (43.42% of the total variance), vertical axis = 3rd factorial axis (14.78% of the total variance). (d) Dispersion of pollen taxa: groups for the percentages of 77 taxa in 86 samples. Horizontal axis = 1st factorial axis (42.44% of the total variance), vertical axis = 3rd factorial axis (31.74% of the total variance)

Laghi dell'Orgials, the reconstructed T_c values are 10 °C higher during the Late Pleniglacial and the Oldest Dryas (OR1) than during the Bölling–Alleröd (OR2) and Younger Dryas (OR3) and no differences are observed between these two periods. At Lago delle Fate, values obtained during the Oldest and Younger Dryas (FA1, 5) are 8 °C higher than during the Bölling–Alleröd (FA2, 3, 4). In all periods, there is a difference of several degrees (2–6 °C) in the values obtained at either site. The standard errors are not shown in Fig. 4, in order to simplify the image, but they do not superimpose.

The warming during the transition between the Lateglacial and the Holocene (OR4, FA6) is similar in both sites (approximately 14 °C). Following this period, the differences between the two sites become increasingly marked, as the two curves

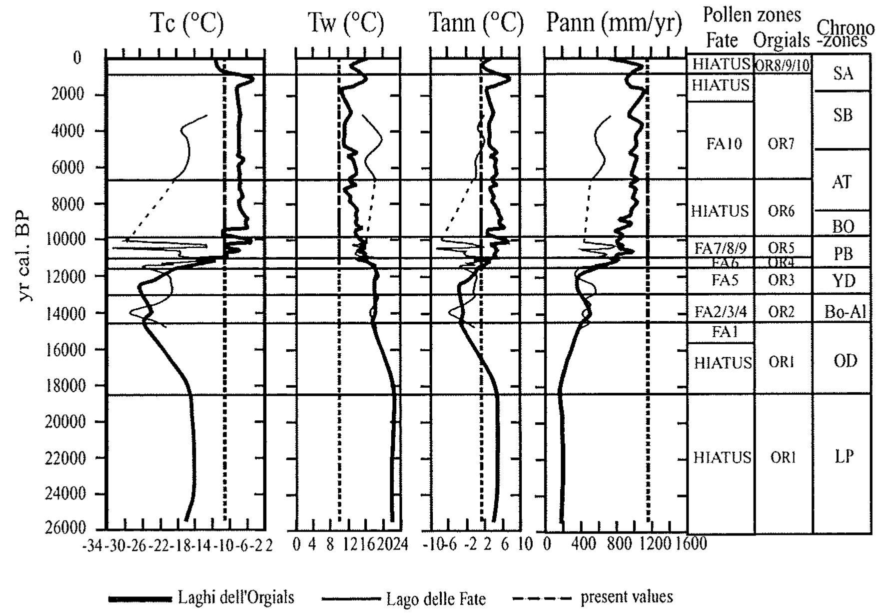

Figure 4 First test for palaeoclimatic reconstruction at the two sites. Reconstructed parameters: T_c: mean temperature of the coldest month; T_w: mean temperature of the warmest month; T_{ann}: mean annual temperature; P_{ann}: mean annual precipitation

show opposite trends between 11 200 and 9950 cal. yr BP (FA6–9, OR4–5). At Laghi dell'Orgials and in the FA6 phase at Lago delle Fate the temperature reconstructions appear to be biased by deciduous arboreal pollen (*Quercus*, *Corylus*, *Betula*) blown uphill from the lowland. The 'best modern analogues' selected are pollen spectra from temperate deciduous forest ($T_c \cong 2\,°C$) and result in excessively warm temperatures (−8 to −2 °C). Following this period, a strong cooling is reconstructed at Lago delle Fate (FA7), with values between −32 and −10 °C, which are generally lower than the Lateglacial values. These abnormally low temperatures, characterised by large fluctuations result from the local development of birch and pine. The 'best modern analogues' are selected from the northern Russian taiga, where coldest month temperatures are around −37 °C. The climatic interpretation of the fir-wood phases (OR6-7 and FA10) gives values of approximately −4 °C at Orgials and about −16 °C at Fate. This difference is due to high *Pinus* percentages in the Fate pollen sequence, arising from the development of a subalpine pine wood around the site (Ortu *et al.*, 2005), and also resulting in the selection of some modern analogues from the northern Russian taiga.

Reconstructed T_w values are particularly high during the Late Pleniglacial and Lateglacial period, ranging between 20 and 22 °C (10 °C more than the present value). The temperatures then decrease progressively from the Lateglacial to the present. The transition to the Holocene (OR4, FA6) is reconstructed as a cooling. The values from the two sites agree, but they decrease further during the following phases (OR5–10, FA7–10).

Trends in T_{ann} values follow those of the coldest month temperatures, which is the limiting factor for mountain vegetation. The annual temperature reconstructed for the Late Pleniglacial and the Oldest Dryas is about 5 °C, which is 3 °C higher than at present. During the transition to the Holocene, the two sites show opposing trends as for T_c.

The P_{ann} variations show an evolution from a dry Lateglacial period to moister phases beginning at the transition to the Holocene. There are some differences in the two curves between 11 200 and 9950 cal. yr BP: P_{ann} vary between 400 and 800 mm at Fate (FA7–9) but between 800 and 1300 mm at Orgials (OR5), with opposing trends for the two sites. The reconstruction of P_{ann} values is affected by the same problem as described above the T_c reconstruction, namely a bias due to a selection of analogues from northern Russian taiga ($P_{ann} \cong 300$ mm).

Tests for palaeoclimatic reconstruction improvement

Modern pollen database improvement

The reconstruction of climate parameters using the improved database (868 modern pollen spectra; Peyron *et al.*, 2005) shows a number of differences with the first analysis (Fig. 5). Coldest month temperatures now show fluctuations during the Lateglacial corresponding to the OR1–3 and FA1–5 phases (Oldest Dryas, Bölling, Older Dryas, Alleröd, Younger Dryas); however, these fluctuations remain inverted, with warm Late Pleniglacial, Oldest and Younger Dryas periods and a cooler Bölling–Alleröd. There continues to be a lack of 'good analogues' during this period, and analogues for the pine-dominated pollen spectra characterising the Bölling–Alleröd are still selected from the northern Russian taiga.

The T_w values are lower (12–18 °C) than in the first reconstruction (12–22 °C), and the curves no longer show a decrease from the Lateglacial to the present. Further, the curves show the expected temperature trends during the Lateglacial, with a 2 °C warming during the Bölling–Alleröd phase and a 4 °C cooling during the Younger Dryas.

The new P_{ann} values are similar for both sequences during the Lateglacial and at the transition to the Holocene. The period between 11 200 and 9950 cal. yr BP is still characterised by opposing trends: at Fate values decrease to 400 mm yr^{-1} with the reduction in *Betula* and the peaks of *Pinus*. This decrease corresponds to that observed in T_c values and result from the selection of modern analogues from the dry northern Russian taiga. This also explains the differences in P_{ann} values of about 400–600 mm yr^{-1} between the two sites for the fir-wood phase after 7400 cal. yr BP (OR7–10, FA10).

Biome constraint

To reduce the bias resulting from the selection of temperate deciduous forest and taiga modern analogues described above,

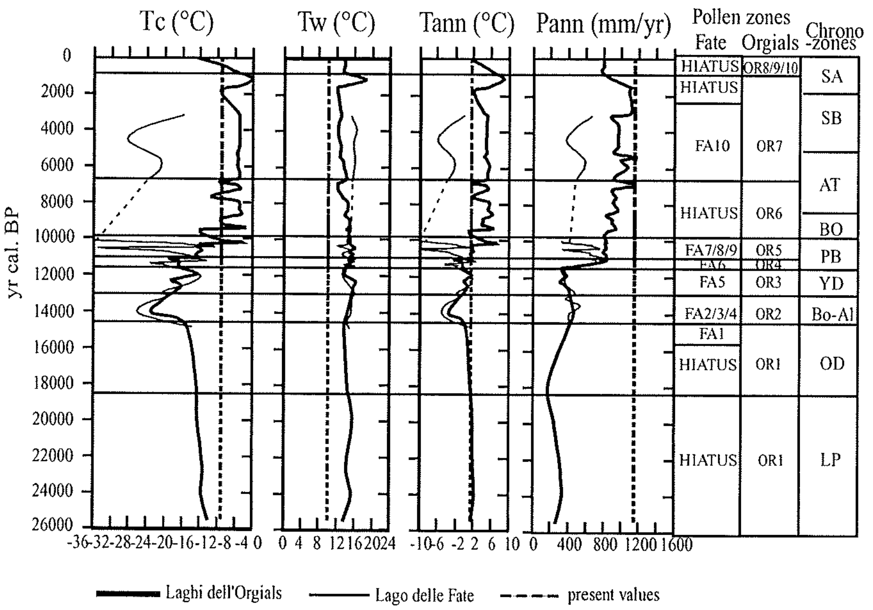

Figure 5 Second test for palaeoclimatic reconstruction at the two sites

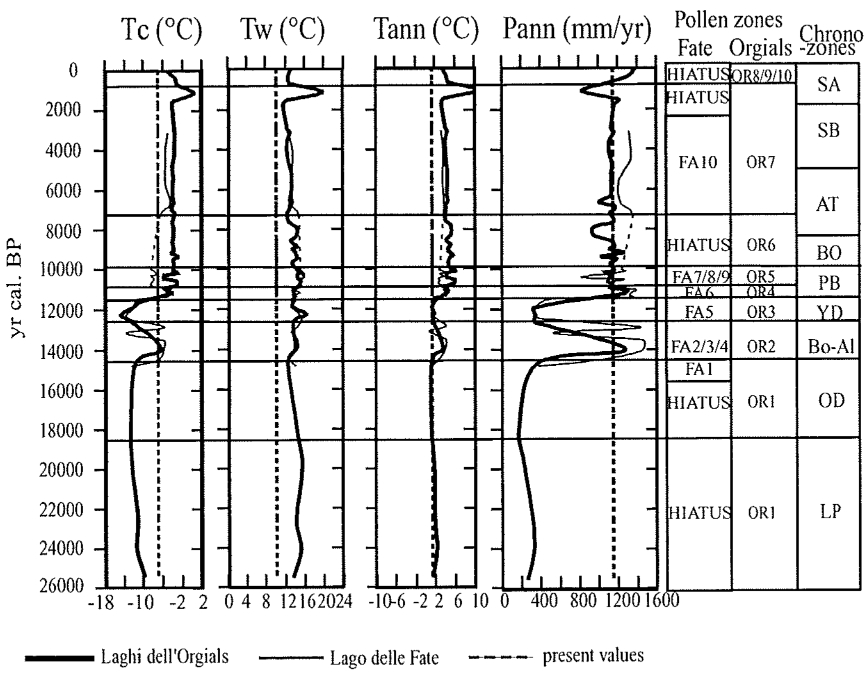

Figure 6 Third test for palaeoclimatic reconstruction at the two sites

a third climate reconstruction was attempted using the biome constraint method (Guiot *et al.*, 1993), in order to exclude modern analogues corresponding to these two biomes.

The results show that both sequences now show the same trend in T_c between the Lateglacial and the Holocene (Fig. 6), and show the expected climatic fluctuations for this period, with lower values for the Late Pleniglacial, Oldest and Younger Dryas than for the Bölling–Alleröd. There is an abrupt rise in T_c values at the beginning of the Holocene, and these then remain at higher than present values throughout the sequence at Laghi dell'Orgials. The difference in values at the two sites of 4–6 °C and the opposing trends (a 4 °C cooling at Fate and a 2 °C warming at Orgials) during the phase between 11 200 and 9950 cal. yr BP still remain.

The T_w curves reconstructed at both sites are now consistent. However, the reconstructed T_w values are between 2 and 6 °C higher than the present-day values during the whole sequence.

The reconstructed T_{ann} in the two sites are in agreement throughout the Lateglacial and Holocene, with the exception of the period between 11 200 and 9950 cal. yr BP. The T_{ann} curves are similar to the T_c ones, with reduced fluctuations. However, T_{ann} values are close to present-day values during the Dryas phases (OR1, 3; FA1, 3, 5), and 4–6 °C higher than today during the Bölling–Alleröd and throughout the Holocene.

The reconstructed P_{ann} values are consistent in the two sites and show trends that follow the expected climatic fluctuations. The amount of precipitation is low (about 200 mm yr^{-1}) during the Late Pleniglacial and the Oldest Dryas, increases during the Bölling–Alleröd period to 1200–1400 mm yr^{-1}, and then falls again to low values (400 mm yr^{-1}) during the Younger Dryas. The beginning of the Holocene is marked by a rise in precipitation values to 1200–1400 mm yr^{-1}, which remain almost constant and close to present values during the Holocene.

The reconstructed precipitation values are in agreement with previous studies (e.g. Lotter *et al.*, 1992; Magny *et al.*, 2001, 2002; Peyron *et al.*, 2005); however, the reconstructed temperatures are abnormally higher than present ones throughout both sequences. This is also in disagreement with previous studies (Heiri *et al.*, 2003, 2004) reconstructing palaeotemperatures close to present ones during the upper part of the Holocene.

Discussion

Analysis of results

The improvement of the modern pollen database and the use of the biome constraint considerably improved the climatic reconstruction, most notably the total annual precipitation. The remaining biases concern mainly the reconstruction of temperature (T_c, T_w, T_{ann}).

Two principal biases remain: (1) the reconstructed temperatures are globally too high when compared to present values; (2) the temperature curves show inverted trends between 11 200 and 9950 cal. yr BP, though the amplitude of these changes has been lower than in the first two reconstruction attempts (Figs 4 and 5). These biases may result from a number of factors.

(1) Temperature is the biological limiting factor in montane regions. In these areas, there is a strong altitudinal temperature gradient (0.55 °C 100 m^{-1}). This results in an altitudinal gradient of vegetation zones that is typical of the Alps from the basal to the alpine belt. This altitudinal vegetation zonation is equivalent to a latitudinal vegetation zonation at a much larger scale (Pignatti, 1979; Ozenda, 1985; Klötzli, 1991). The uphill transport of pollen by wind mixes together pollen from different vegetation belts, and therefore different parts of the temperature gradient. Lakes and peat bogs situated at high elevation may receive non-local pollen from all belts situated at lower altitude. During reconstruction, modern analogues may be selected from lower elevations, causing an overestimation of palaeotemperature.

The uphill transport of pollen by wind to subalpine and alpine zones was first recorded in the European Alps (Rudolph and Firbas, 1926; Barthélémy and Jolly, 1989; David, 1993a, 1997; Brugiapaglia *et al.*, 1998; Ortu, 2002) and has also been

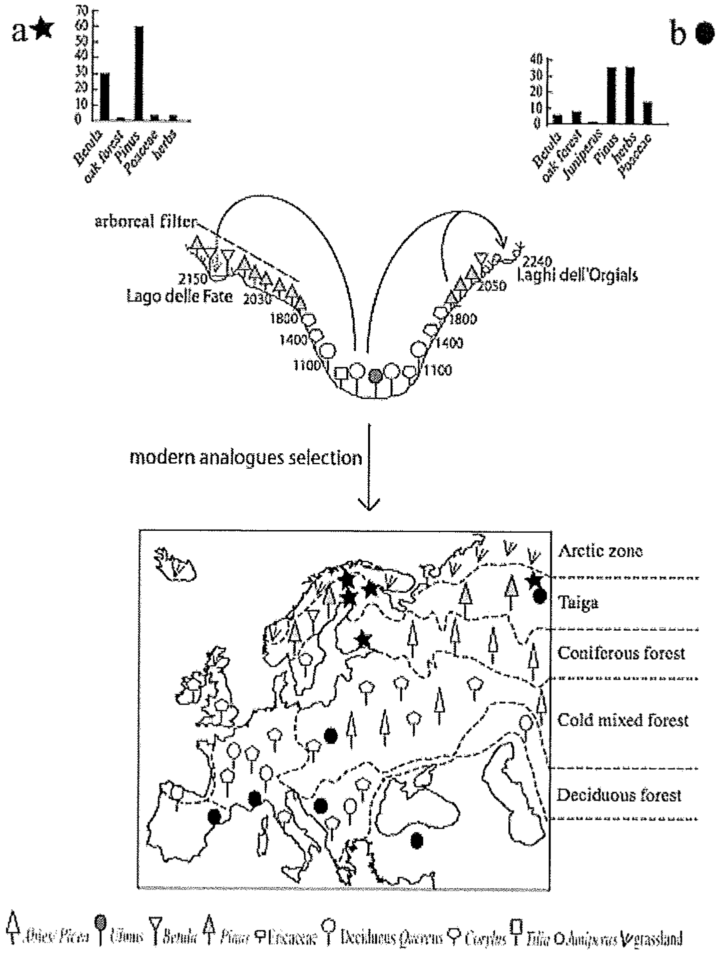

Figure 7 Pollen capture mechanisms in high elevation wooded (a) and non-wooded (b) sites. In this example, this results in the selection of modern analogues in different European biomes for fossil samples recording the same period (ca. 10 600 cal. yr BP) at the site (a) Lago delle Fate and (b) Laghi dell'Orgials

found in the mountains of Africa, South America, New Guinea (Flenley, 1973) and New Zealand (Moar, 1970; Randall, 1990a, b). This phenomenon becomes more important when the arboreal cover decreases (Fig. 7), as closed tree formations have a filtering action on non-local pollen transported by wind, decreasing relative percentage. The modern pollen database lacks a sufficient number of high-elevation pollen spectra. This yields two possible consequences depending on whether the analysed fossil sample comes from a wooded (a) or a non-wooded (b) phase (Fig. 7):

(a) If a fossil pollen sample reflects a wooded phase (Fig. 7(a)), the pollen-based climate reconstruction is generally less affected by non-local contributions. If the modern database lacks modern pollen spectra from mountain areas, analogues are chosen from vegetation formations with a latitudinal climatic range that is equivalent to the altitudinal range, which is represented by the fossil sample. However, the subalpine pine wood at Fate has no modern equivalent in the modern pollen databases, because of the anthropogenic suppression of subalpine *Pinus cembra* formations in the Alps. These formations only exist today in a small number of protected areas, e.g. Bosco de l'Aleve, southwestern Italian Alps. A floristically comparable formation can be presently found in *Pinus sylvestris* woods, which has, however, a larger climatic range. This results in the selection of modern analogues from the northern Europe taiga. The very low temperature and precipitation values obtained for the Fate sequence are due to a mixture of Scandinavian analogues (vegetal formations climatically equivalent to the alpine ones) and Russian taiga analogues (too cold, too dry). This led us to constrain the analogues selection by suppressing the taiga biome. However, this constraint represents a loss of information as the subalpine pine wood is ecologically and climatically close to the Scandinavian taiga open coniferous forest (Pignatti, 1979). Further improvement of the modern pollen database by the addition of modern pollen spectra from subalpine and boreal pine wood (collected in selected areas preserved from anthropogenic impact) could help solve this problem, providing a larger number of modern analogues for a correct quantitative interpretation.

(b) When the fossil pollen sample reflects a phase above the upper forest limit (Fig. 7(b)), the reconstruction may be biased by the absence of modern pollen spectra from the subalpine and the alpine belts, which are characterised by pollen assemblages dominated by a blend of non-local taxa coming from different vegetation belts in the modern pollen database. This may result in the selection of modern analogues from a number of different vegetation formations, with potentially quite different climatic ranges (Fig. 7(b)). At Laghi dell'Orgials, the presence of *Pinus* in fossil samples due to regional wind-driven pine pollen transport led to a selection of modern analogues from northern Europe taiga and an underestimation of temperatures; for the same period, the presence of oak forest pollen blown uphill by wind from the lowland led to a selection of analogues corresponding to the temperate deciduous forest biome and an overestimation of temperatures. When a reconstruction was attempted with a constraint to avoid these

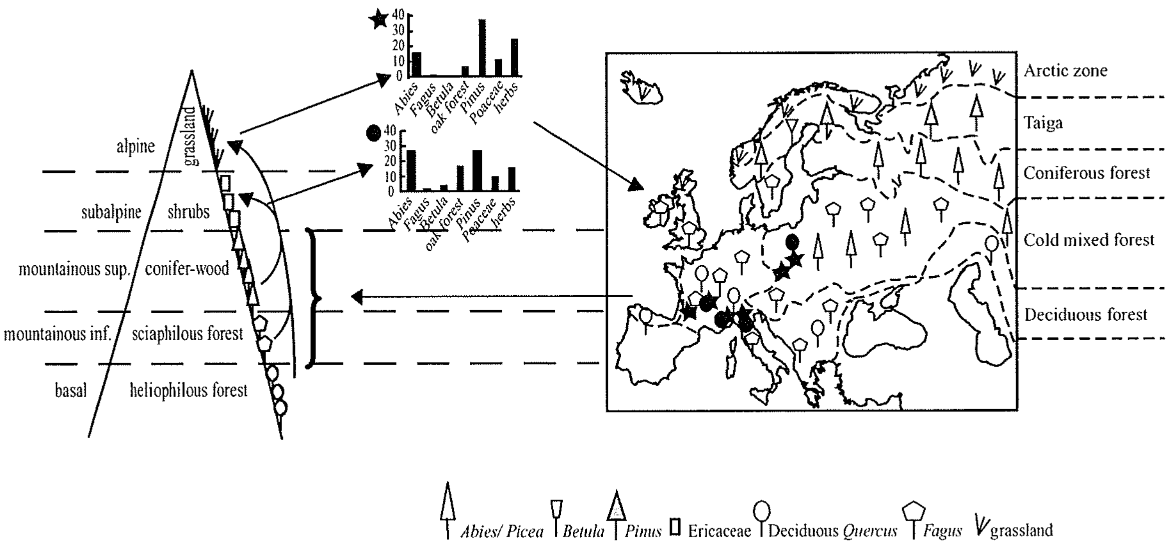

Figure 8 The dominance of coniferous pollen (mainly *Pinus*) in pollen spectra from the subalpine and the alpine belts causes the selection of modern analogues from cold mixed forests, which climatically correspond to the montane belt

biomes, the temperatures remained too warm. This may be due to the selection of an incorrect biome or late Holocene human impact.

An incorrect selection of modern spectra belonging to the biome 'cold mixed forest' rather than alpine pastures may occur, due to the dominance of conifer pollen and the presence of deciduous taxa pollen in fossil spectra. This biome corresponds climatically to the montane belt (Fig. 8), with higher temperatures. This is mainly due to a lack of modern alpine and subalpine pollen spectra in the databases and to the fact that the herbaceous taxa are not dominant in pollen spectra from sites situated over the timberline (Barthélémy and Jolly, 1989; David, 1993a, 1997).

Alternatively, human activities during the late Holocene lowered the timberline (David, 1993a; Wick, 1994; Ortu *et al.*, 2003). This means that modern alpine pollen spectra represent vegetation formations that should be at higher elevation (Fig. 9). As this has no influence on the climatic gradient but instead on the vegetation zonation and its upper limit, modern pollen spectra from pastures located above the present timberline are situated at lower altitudes than under equivalent climates in the past and are associated with warmer temperatures (Fig. 9).

(2) From approximately 11 200 to 9950 cal. yr BP, a number of differences are observed between the two sequences in both the temperature and precipitation curves. During this period, the Fate sequence records a local pollen signal, while the sequence at Orgials is dominated by non-local pollen. The temperature estimations at Laghi dell'Orgials (Fig. 6) for this period are therefore overestimated owing to the transport of pollen from lower vegetation belts (Fig. 7). This interpretation is supported by the coherence of the temperature curves for the pollen zones FA6 and OR4, which are dominated by non-local pollen in both sites, prior to the development of birch at Fate (Fig. 2: pollen zones FA6 and OR4). Further, the fluctuations in precipitation and average annual temperature are in opposition between the two sites at this time. A closer examination shows that the reconstructed climate fluctuations at Fate follow the variations in *Betula* percentages, which may have non-climatic causes. For example, the peak of *Betula* at 10 400 cal. yr BP (Fig. 2), may be due to a local event such as a fire (Ortu, 2002).

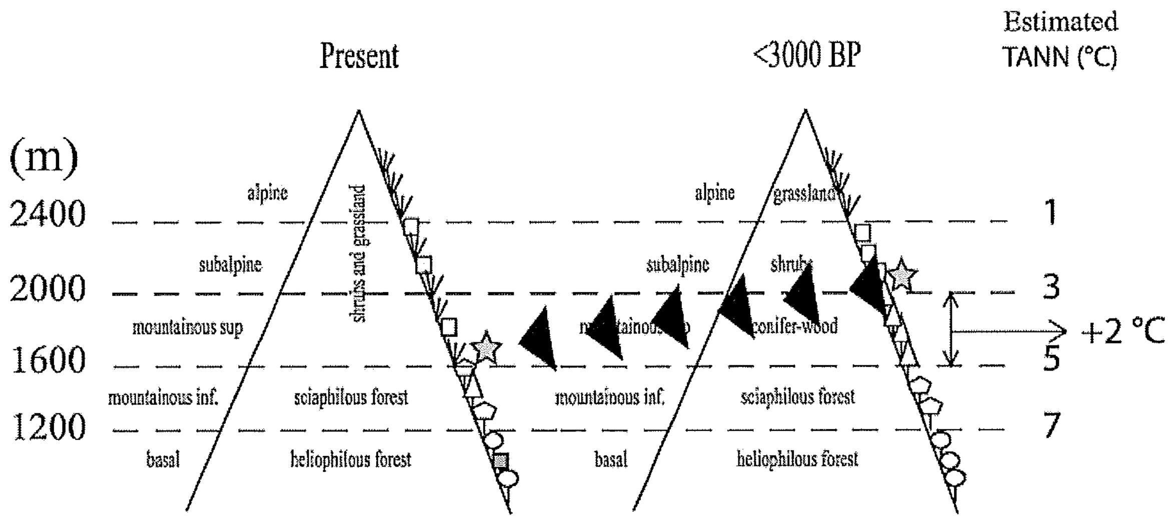

Figure 9 Mechanisms of modern analogues selection in mountain areas disturbed by human activities

Perspectives for improvement

On the basis of our study, some possible solutions to improve the reliability of palaeoclimatic reconstructions in mountain areas may be proposed.

The modern pollen database should be improved by the addition of modern pollen spectra from mountainous areas that are characterised by natural or pseudo-natural vegetation, reflecting climate conditions. Some vegetation refugia still exist in protected areas, in which the original vegetation can be found. Some of these areas in the Alps have been sampled and the analyses are in progress. However, samples taken from pseudo-natural vegetation developed in a vegetation belt at high elevation will still be affected by pollen transported from disturbed lower vegetation belts, especially above the timberline (Fig. 7(b)). Human impact on the natural oak forest in the lowland has facilitated the introduction of allochthonous taxa, e.g. the replacement of deciduous oak by chestnut. This gives rise to a pollen signal that can be quite different from the naturally occurring signal. To help improve the reconstruction technique, it may be possible to use the presence of these 'foreign' taxa to quantify the anthropogenic disturbance in the lowland vegetation and its representation in pollen spectra at different altitudes.

By relating the modern pollen spectra to the vegetation belt from which they were taken, it may be possible to develop a different approach to mountain palaeoclimatic reconstructions. The correspondence analysis shows that fossil pollen samples recording the same vegetation phase can be grouped (Fig. 3) despite the differences in assemblages. Knowing the vegetation belt that corresponds to a fossil sample would allow the vegetation zonation to be reconstructed and climatic variations to be inferred. A possibility would be to correct the climate attributed to modern pollen samples, so that it corresponds to the climatic range of the vegetation type recorded in the sample, rather than the site location. This would still require careful interpretation of the local/regional aspect of the analogue climate, but should help to reduce the biases caused by anthropogenic impact.

Conclusion

The comparison of results obtained from two high-elevation pollen sequences has highlighted the problems of quantitative reconstruction of climatic parameters. The differences in the temperature and precipitation values and in the reconstructed trends are mainly due to differences in pollen records related to local conditions at the two sites (altitude, aspect, geomorphology, vegetal cover). The main differences result from the dominance of local or non-local pollen in the samples. However, correspondence analysis shows that fossil samples recording the same vegetation phase can be grouped, despite these local differences. Analysis of the reconstructed temperatures shows that biases are mainly due to a poor choice of modern analogues caused by a lack of high-elevation pollen spectra in the modern database. This bias was partially reduced by constraining the analogue choice to exclude modern samples corresponding to temperate deciduous forest or taiga biomes. The new results show consistent trends of climate changes, which are largely coherent in both sites. The reconstructed temperature values remain too high. We attribute this to the presence of pollen from different vegetation belts, characteristic of pollen spectra from high-elevation sites, in particular above the timberline. The bias towards overestimated temperatures may be further reduced by the addition of more pollen spectra from the subalpine and the alpine belts. We underline the need for more precise selection strategies when using modern samples for climate reconstruction in high-altitude mountain sites, including the need to collect new samples from selected refugia areas to reduce the bias caused by anthropogenic transformation of the landscape, and an analysis of the relationship between pollen spectra and vegetation zonation. The elaboration of a new methodology, specific to mountainous areas, may be necessary, reconstructing the altitudinal gradient of palaeovegetation as a first step and subsequently inferring climate variations from changes in this zonation.

Acknowledgements This study has been financially supported by the French CNRS within the framework of the ECLIPSE Program (Past Environments and Climates). SB was supported by the EU project MOTIF 'Models and Observations to Test Climate Feedbacks' (EVK2-CT-2002-00153).

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Lateglacial and Holocene climate oscillations in the South-western Alps: an attempt at quantitative reconstruction.

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Abstract:

The high topographic complexity of the Alpine region is at the origin of important climate differences which characterise the different areas of the Alps. These differences might have had a strong influence on vegetation and on migrations of human populations in the past. Based on an improved data base containing about 3000 modern pollen samples, the standard "Modern Analogue Technique" has been applied to five pollen sequences from the subalpine belt of the South-western Italian Alps (Laghi dell'Orgials, 2240 m, Lago delle Fate, 2130 m, Torbiera del Biecai, 1920 m, Rifugio Mondovì, 1760 m, Pian Marchisio, 1624 m) to provide quantitative climate estimates for the Lateglacial and Holocene periods. Consistent climate trends are reconstructed for the different sequences. The sensitivity of vegetation to climate variations changed at various sites with the fluctuations in the boundaries of vegetation belts. Sites above the tree line recorded lower temperature values and less important variations. Sites recorded in detail the climate variations when they were located at the limit of two ecotones. No perfect analogues were found for the Oldest and Younger Dryas, when climate was supposed to be cold and dry. Climate was close to present-day values during the Bølling/Allerød interstadial. At the beginning of the Holocene, climate changed to warmer and moister conditions; a high number of climate fluctuations are recorded at several sites. A climate optimum is recorded at the Atlantic period, which caused a development of fir above its present-day altitudinal distribution. Climatic differences recorded at the various sites are discussed taking into account the limits of the method.

Keywords: Alps, Climate, Holocene, Lateglacial, Modern Analogue Technique.

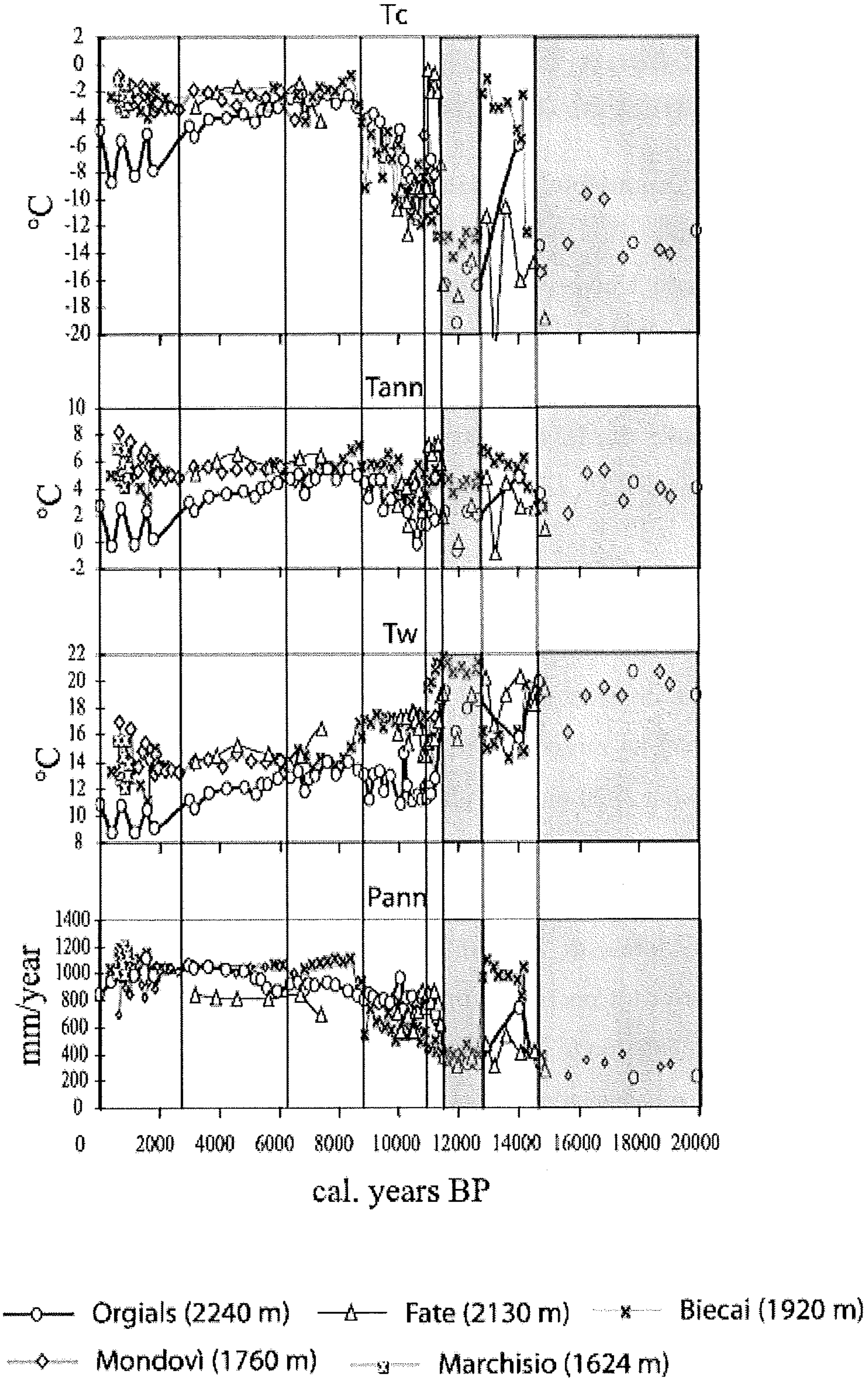


Figure 5: Palaeoclimate reconstruction at the five sites. Reconstructed parameters: T_c: mean temperature of the coldest month; T_w: mean temperature of the warmest month; T_{ann}: mean annual temperature; P_{ann}: mean annual precipitation.

Thursday 20 september

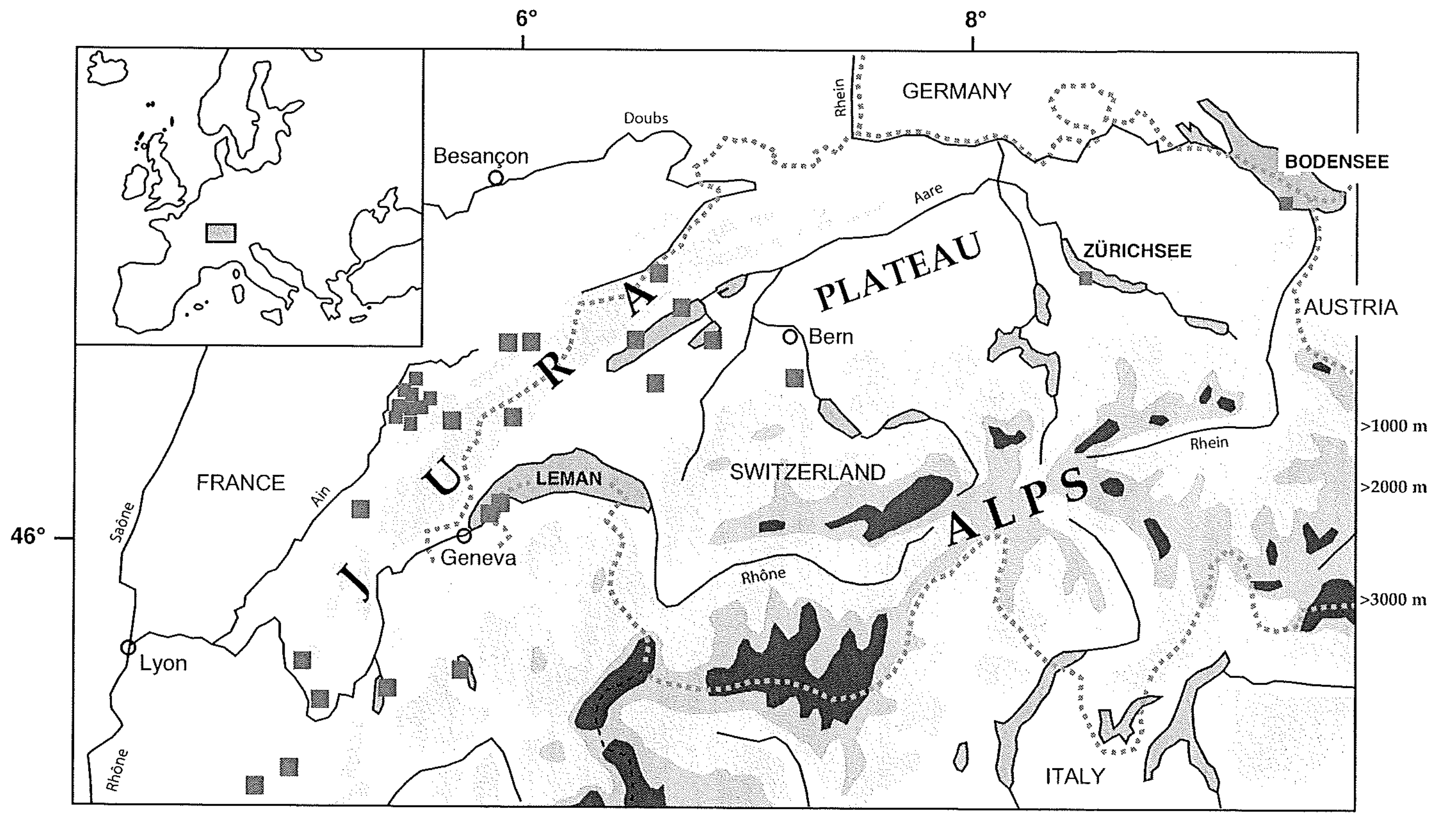
Lakes Maclu
Lake Chalain

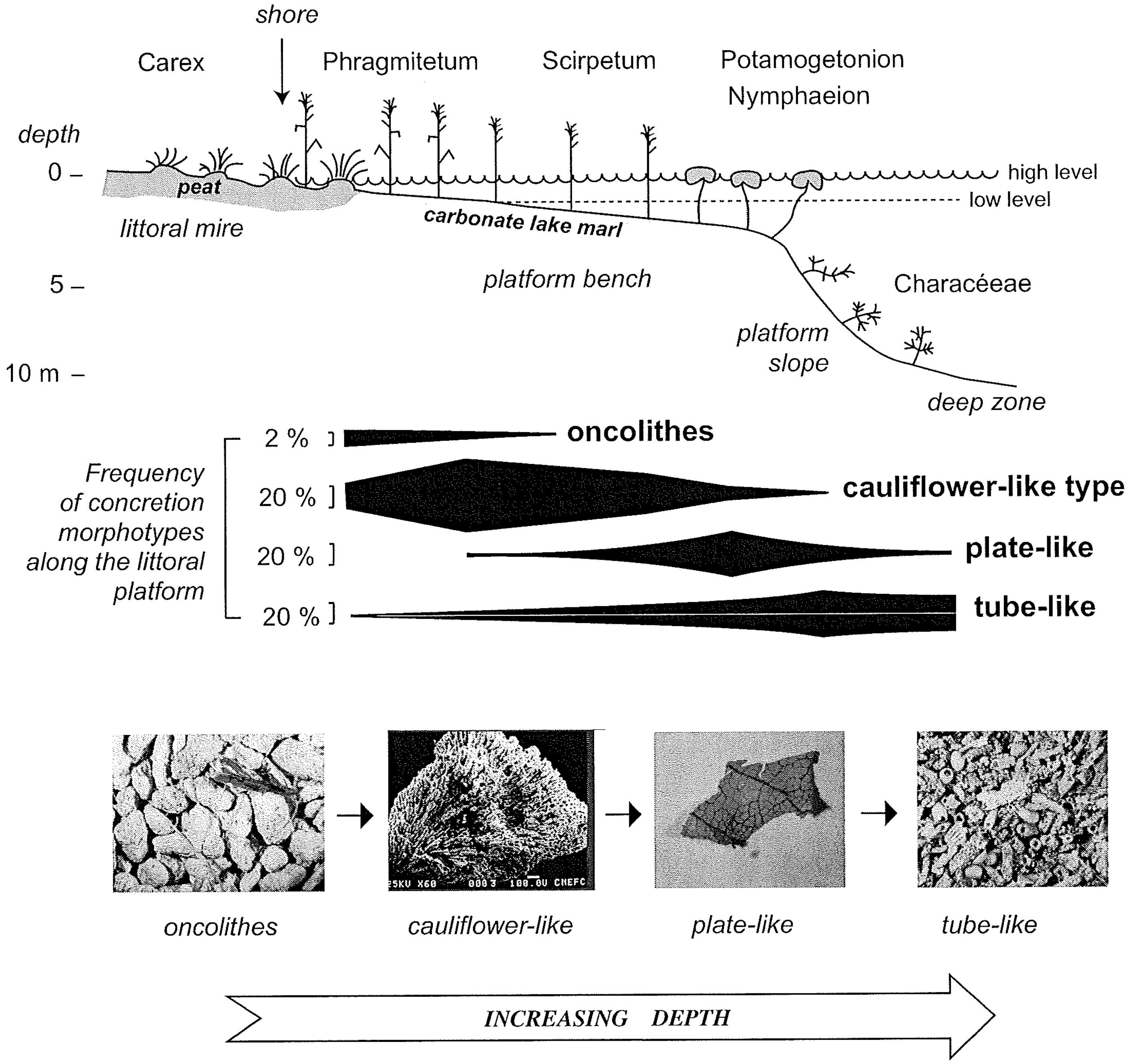
Holocene fluctuations of lake levels in west-central Europe: methods of reconstruction, regional pattern, palaeoclimatic significance and forcing factors

Michel Magny[1A]

Abstract

A detailed lake-level record for the Holocene period has been derived from sediment sequences from 26 lakes in west-central Europe plus a data set of ca 200 radiocarbon, tree-ring and archaeological dates. The data suggest a highly variable Holocene climate characterised by a succession of 15 centennial-scale phases of higher lake level. A comparison of palaeohydrological data collected in western Europe suggests contrasting patterns of hydrological changes in response to Holocene climate cooling phases with wetter conditions over the mid-European latitudes, whereas northern and southern latitudes experienced a drier climate. Quantitative estimates obtained from a method combining pollen and lake-level data indicate that phases of higher lake level coincided with an increase in annual precipitation, a decrease in summer temperature and a shortening of the growing season. Finally, a comparison of the mid-European lake-level record with the atmospheric ^{14}C record, the GISP2 Polar Circulation Index record, the North Atlantic Ice-Rafted Debris record and meltwater outbursts in North America and Europe supports the hypothesis that variations in solar activity were a major driving factor of Holocene climate oscillations.

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5 –
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peat
littoral mire
high level
low level
carbonate lake marl
platform bench
Characéeae
platform
slope
deep zone
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of concretion
morphotypes
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platform
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20 %
20 %
20 %
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plate-like
tube-like
oncolithes
cauliflower-like
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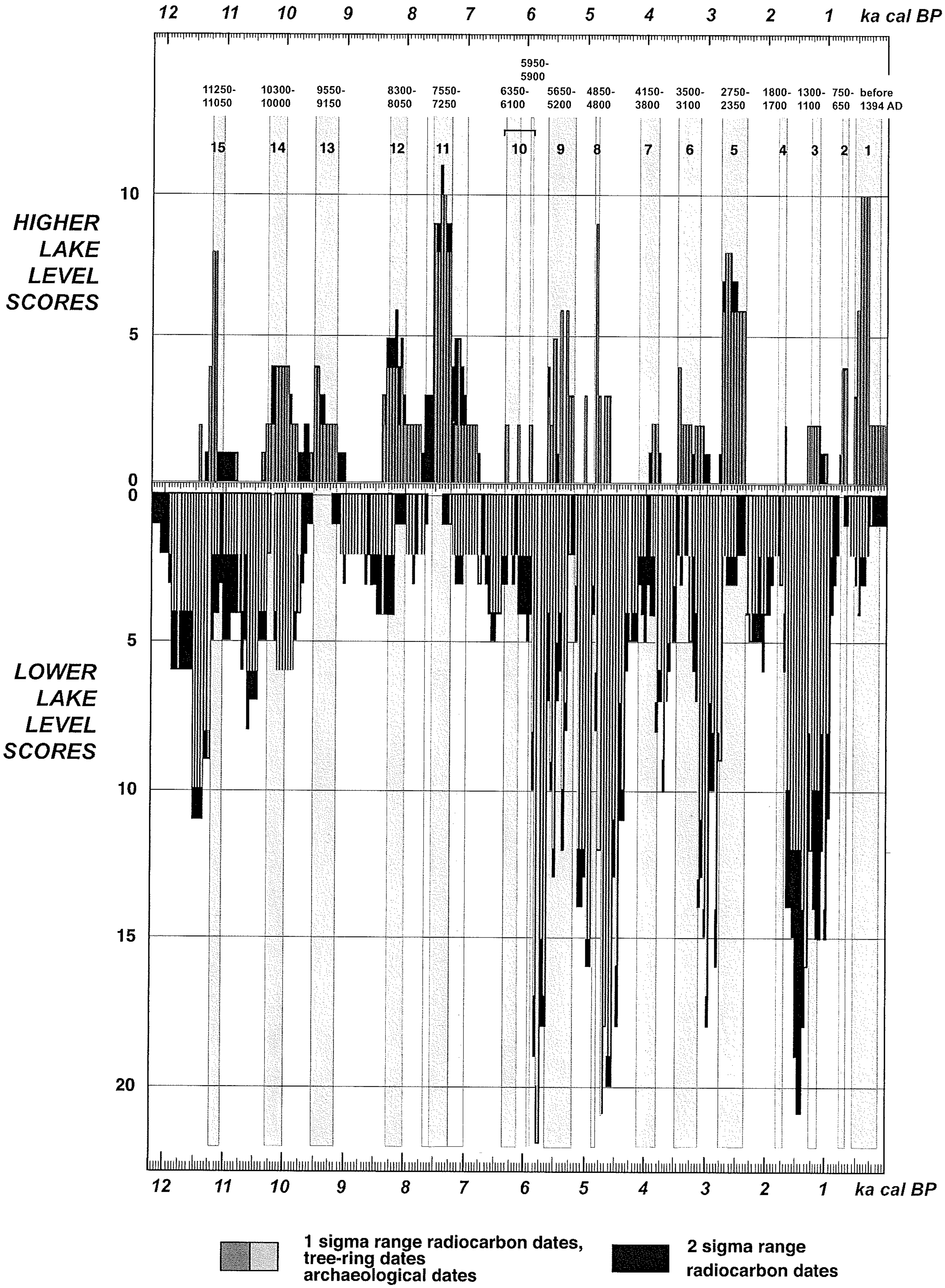

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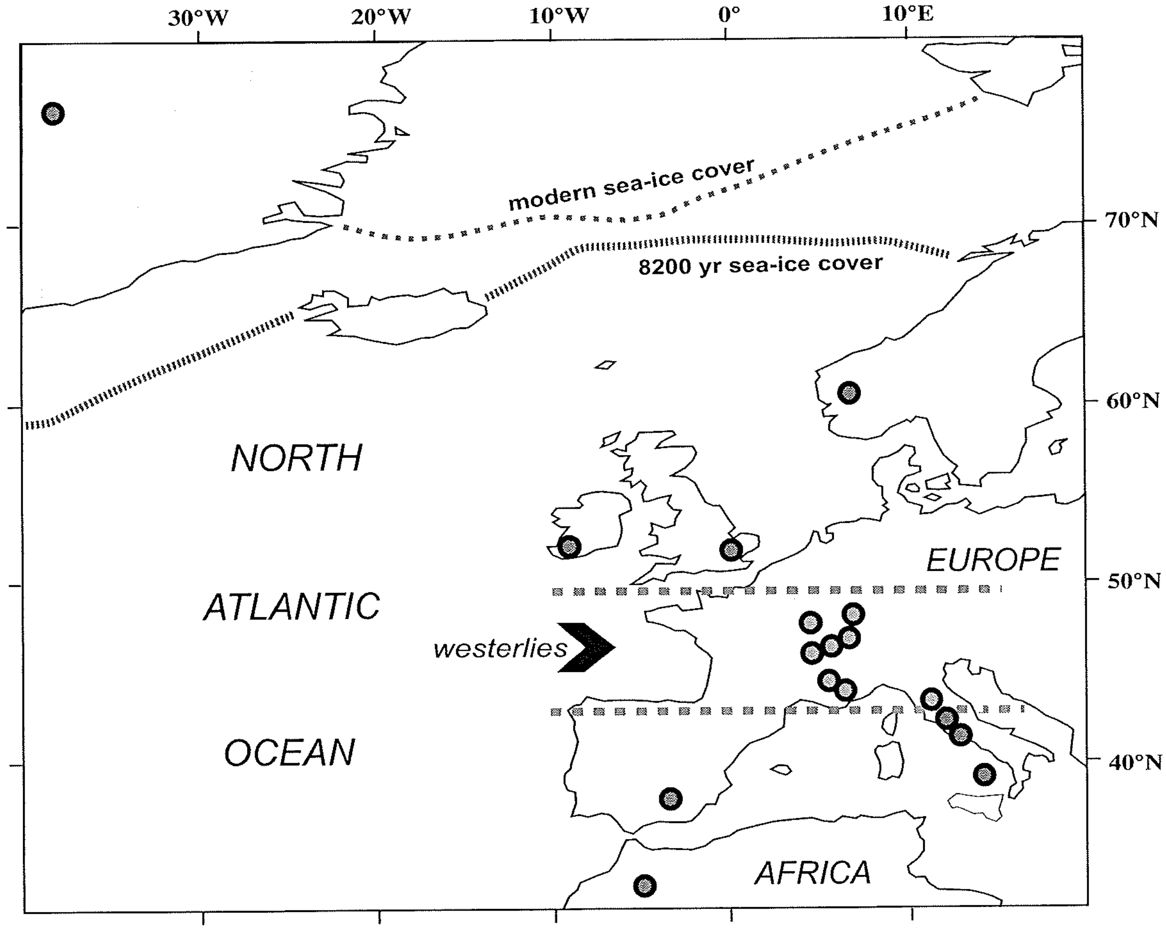

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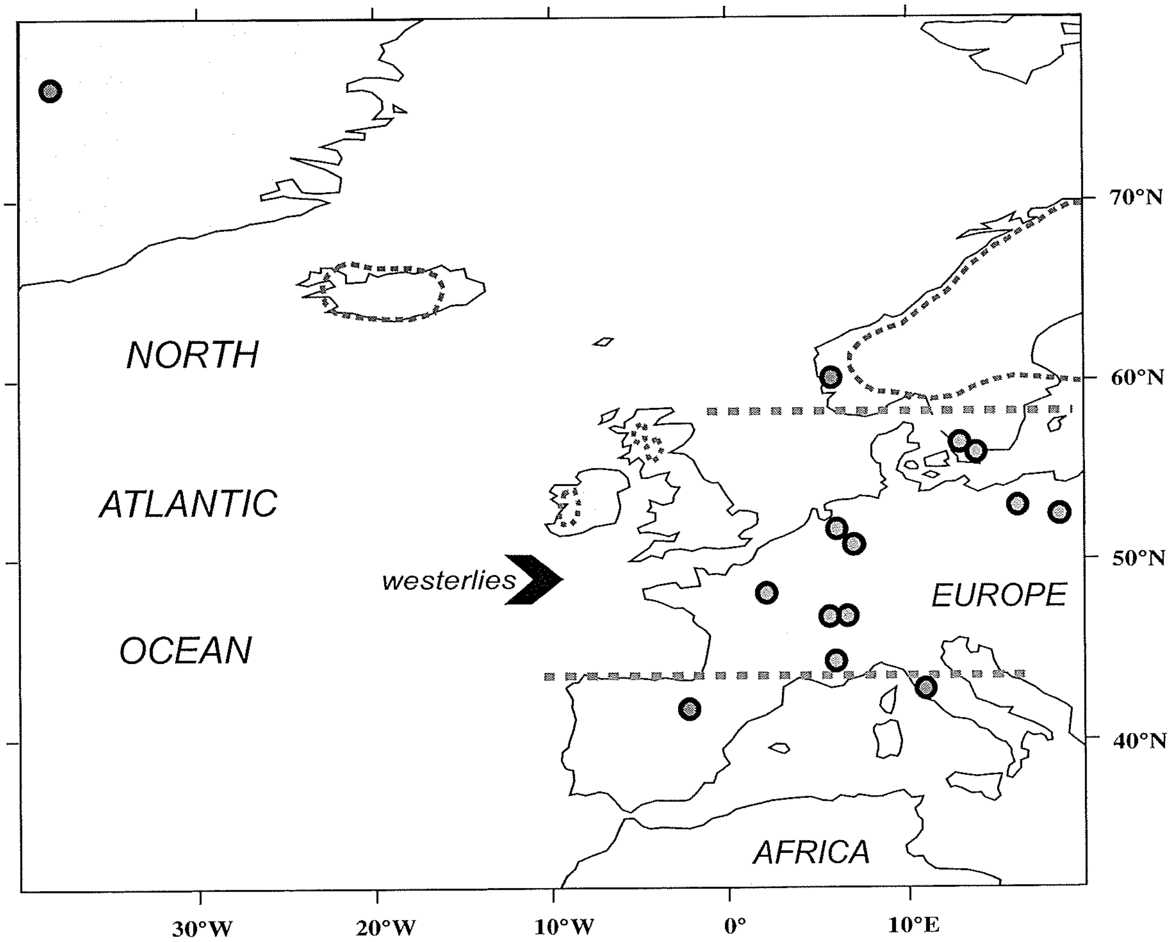

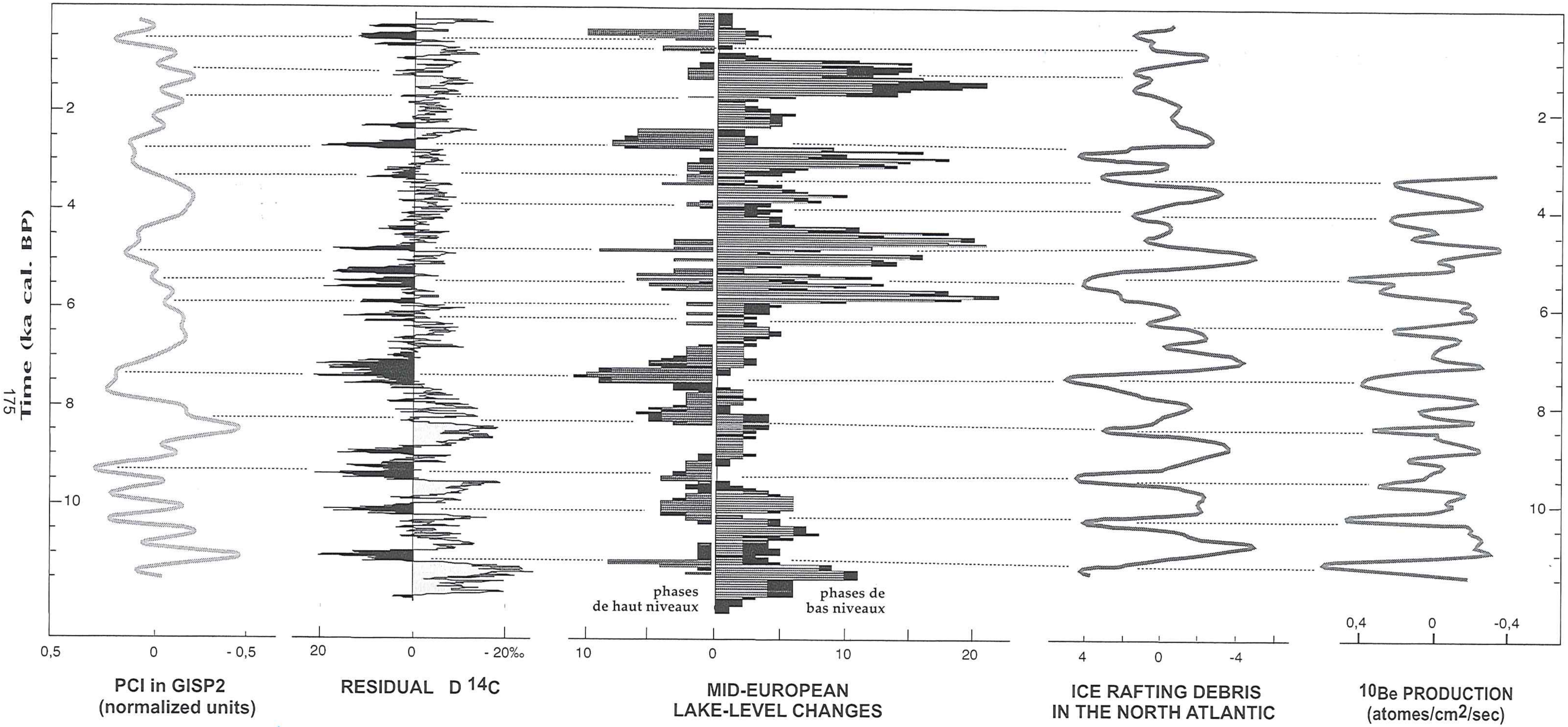

Polar Circulation
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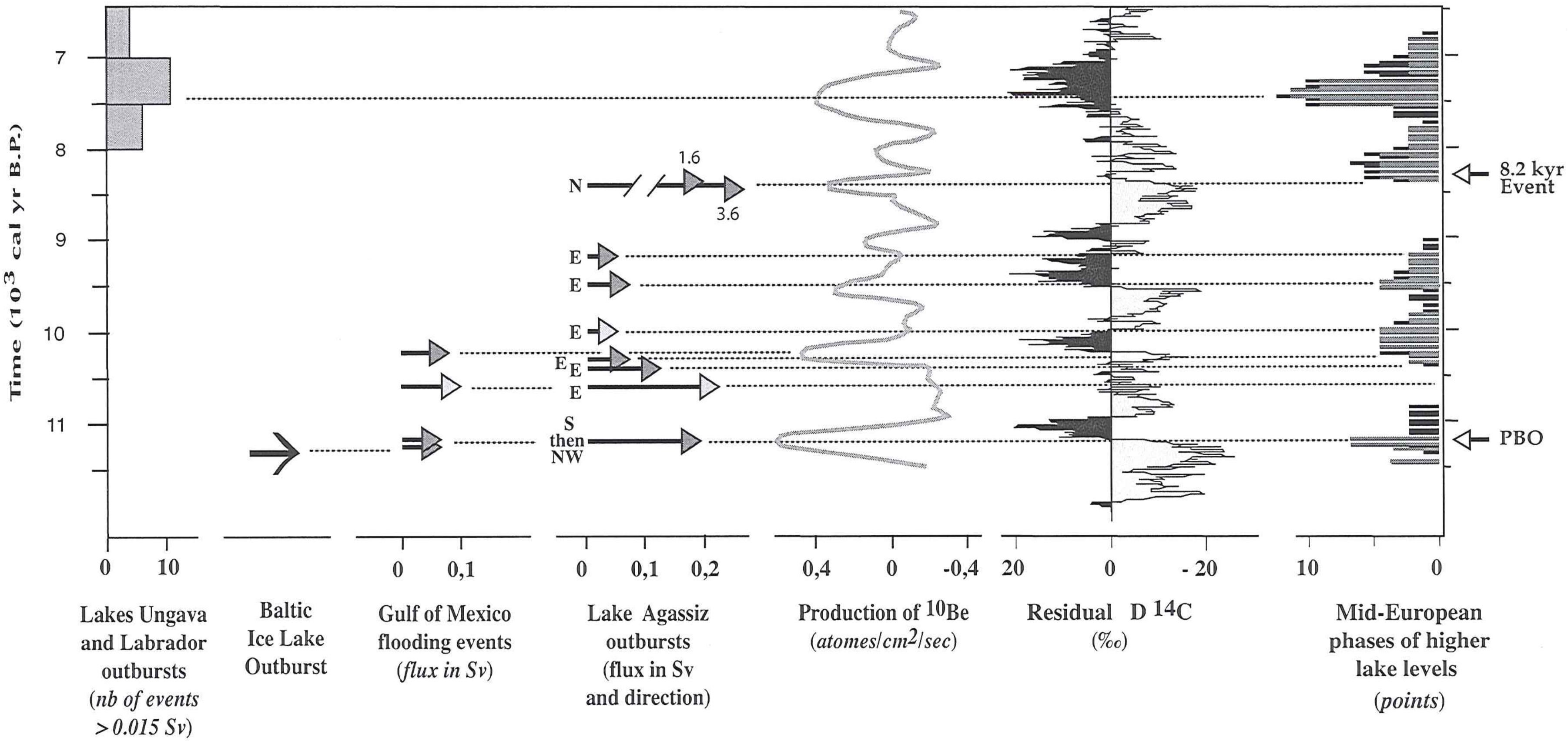
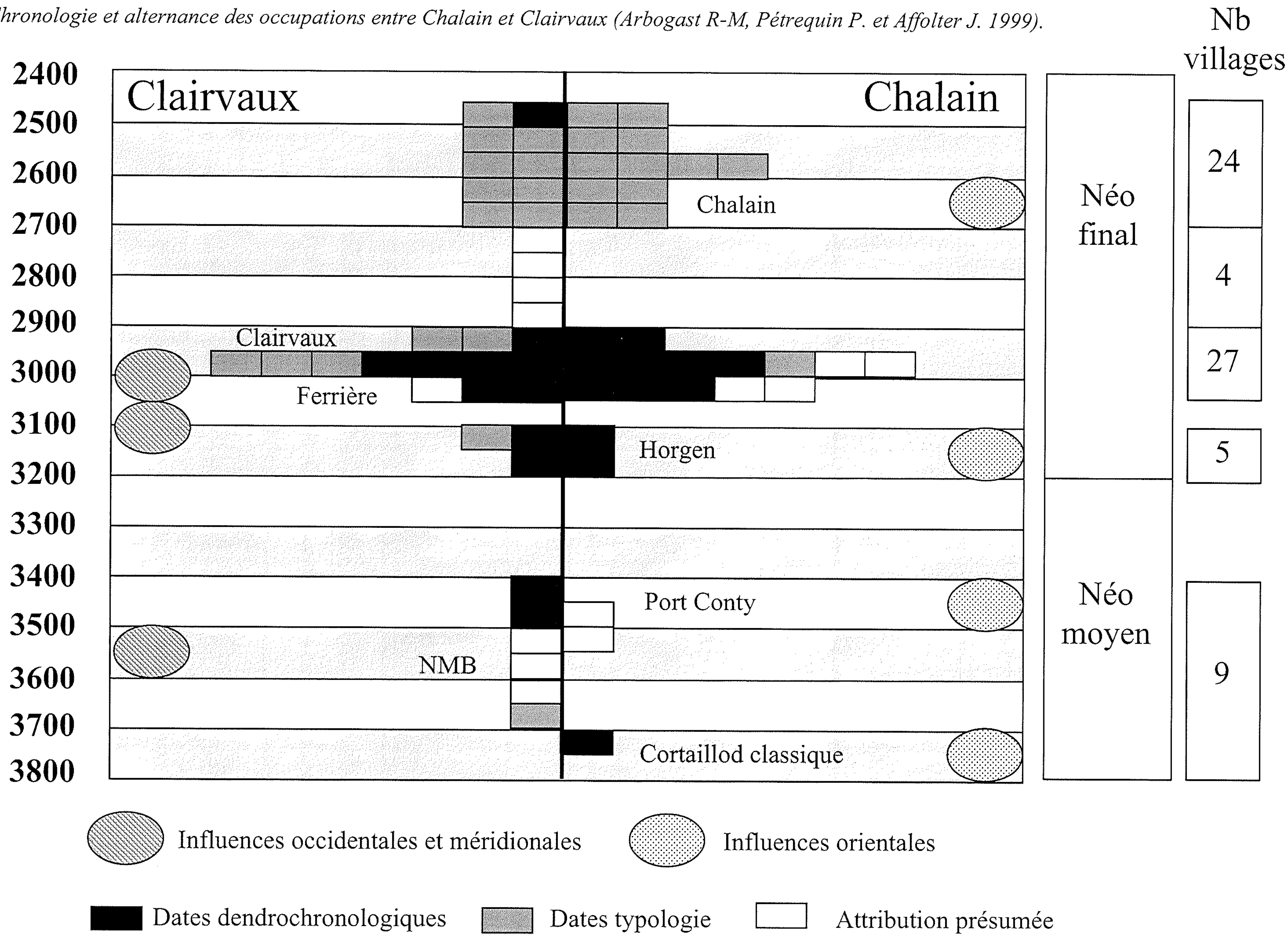
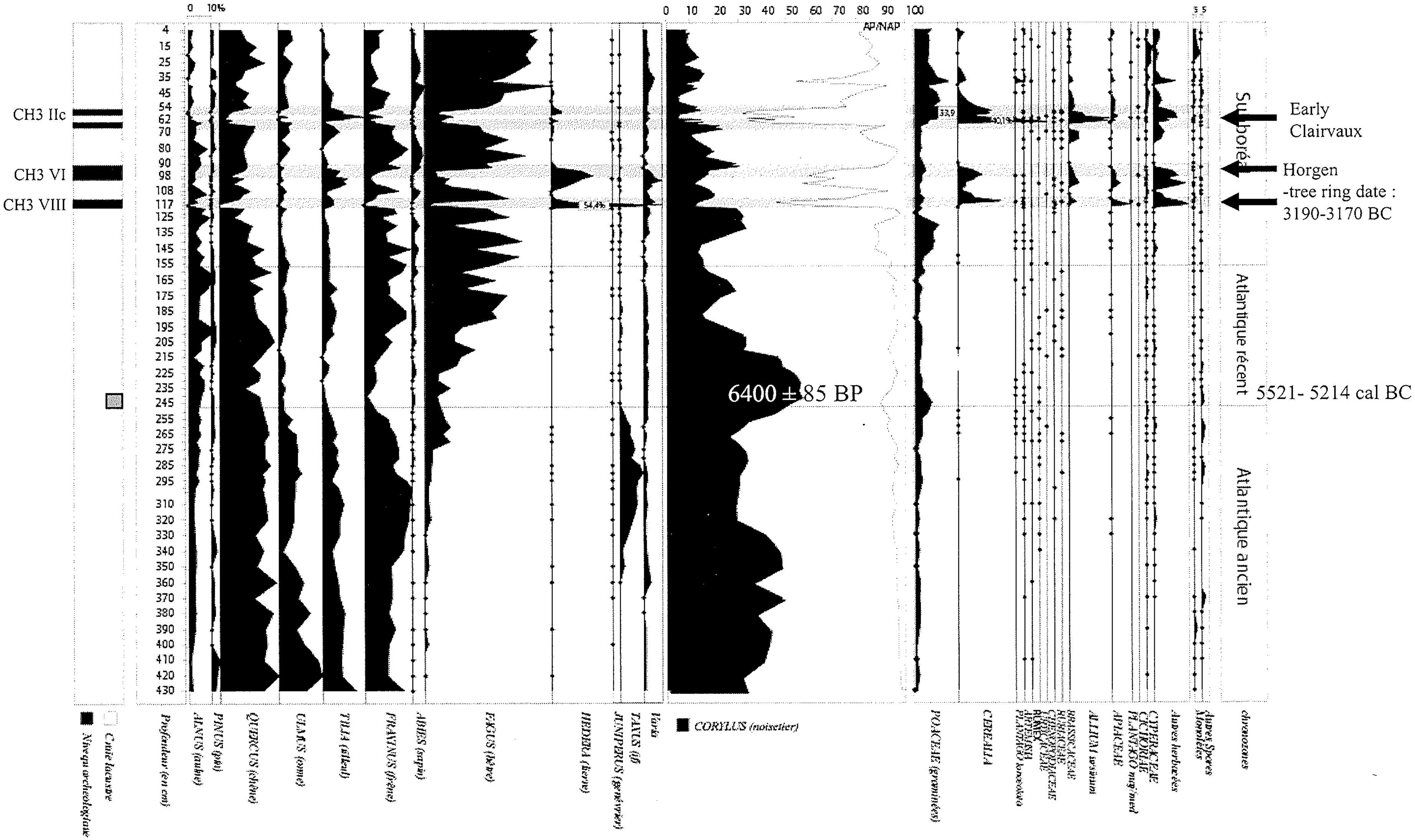


Encyclopedia Quat. Sc.

Chronologie et alternance des occupations entre Chalain et Clairvaux (Arbogast R-M, Pétrequin P. et Affolter J. 1999).



Chalain, station 3 - Richard H., 1986



Quantitative reconstruction of climatic variations during the Bronze and early Iron Ages based on pollen and lake-level data in the NW Alps, France

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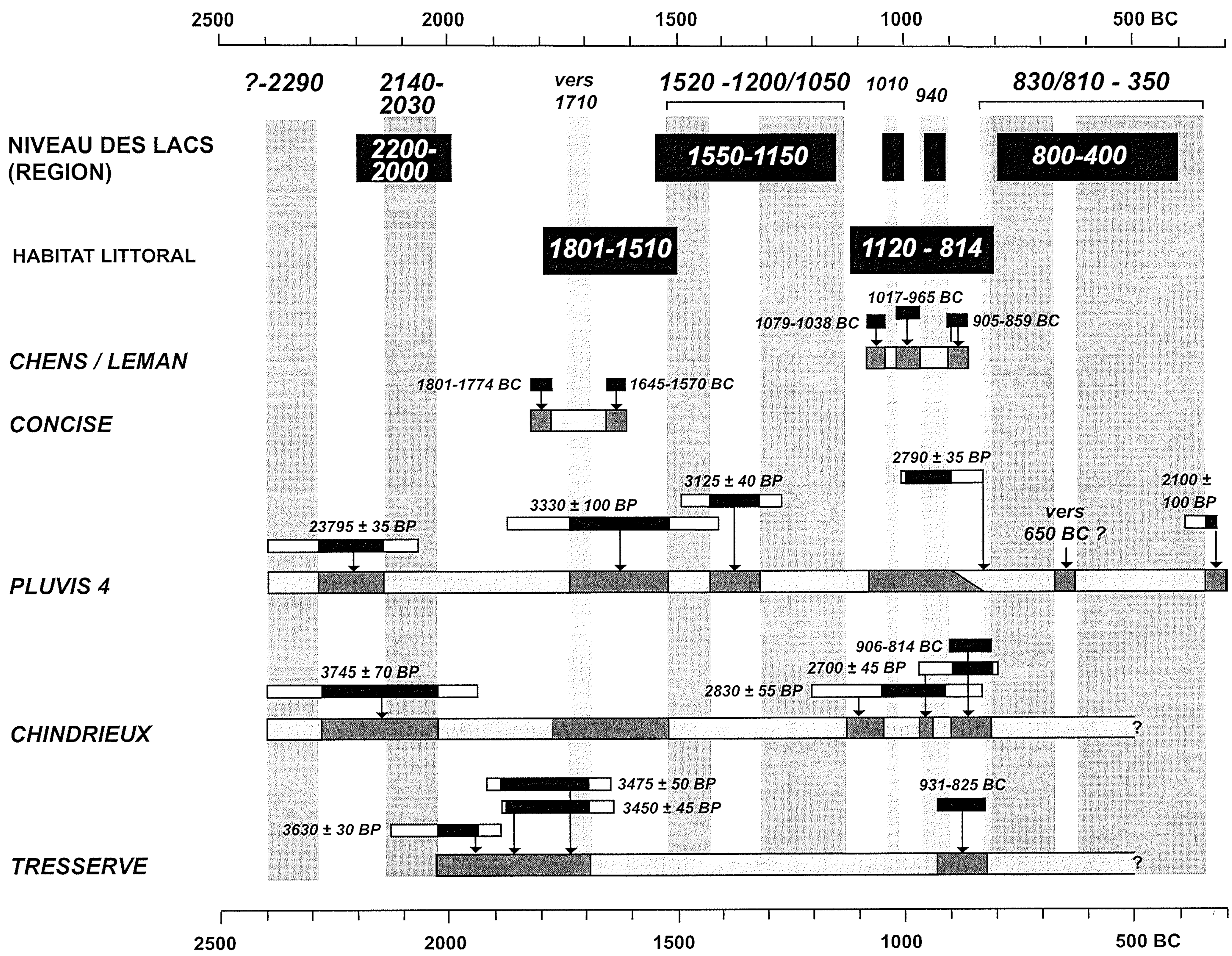
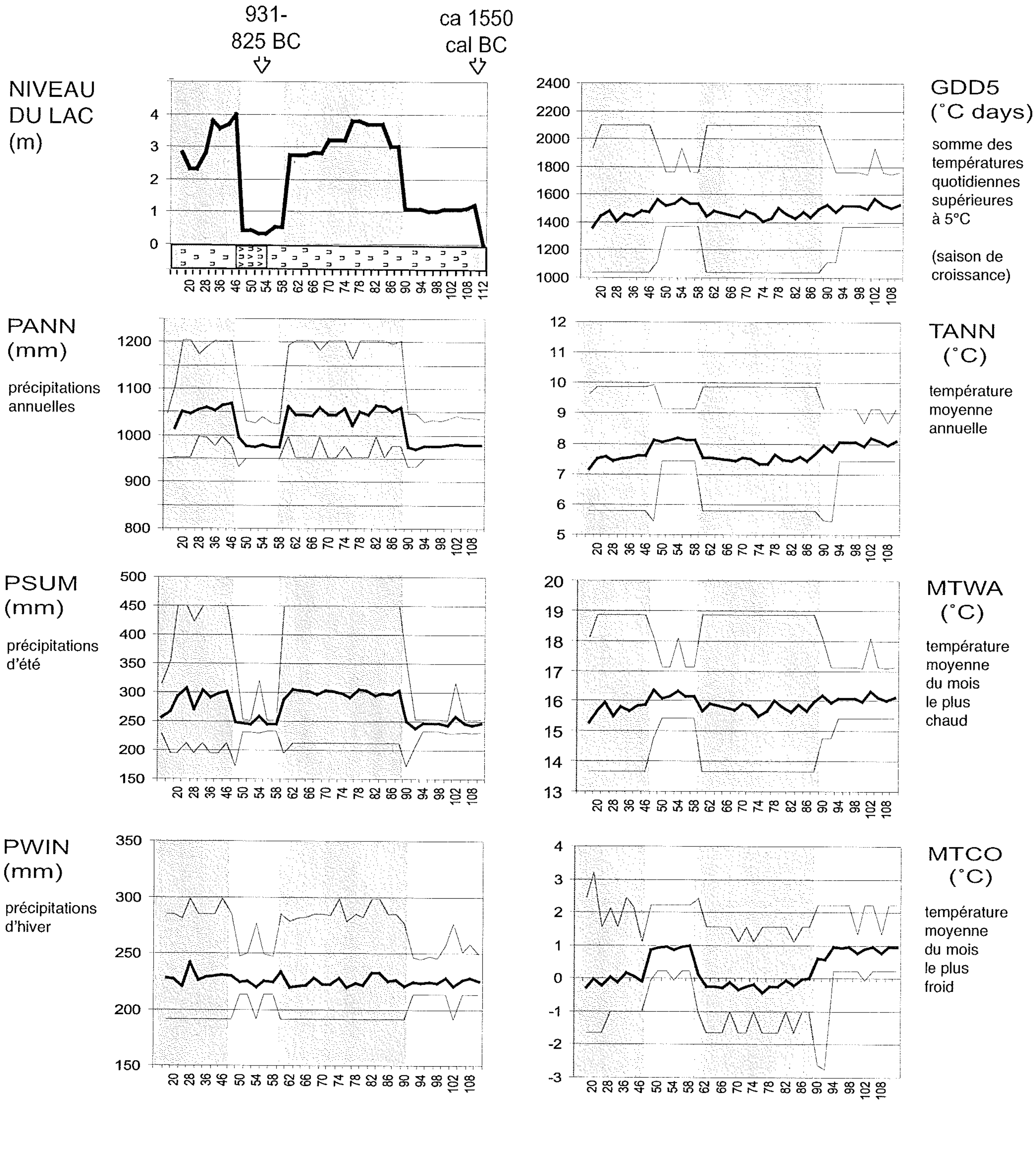
Abstract

Vegetation and lake-level data from the archaeological site of Tresserve, on the eastern shore of Lake Le Bourget (Savoie, France), are used to provide quantitative estimates of climatic variables over the period 4000-2300 cal BP in the northern French Pre-Alps, and to examine the possible impact of climatic changes on societies of the Bronze and early Iron Ages. The results obtained indicate that phases of higher lake level at 3500-3100 and 2750-2350 cal BP coincided with major climate reversals in the North Atlantic area. In west-central Europe, they were marked by cooler and wetter conditions. These two successive events may have affected ancient agricultural communities in west-central Europe by provoking harvest failures, more particularly due to increasing precipitation during the growing season. However, archaeological data in the region of Franche-Comté (Jura Mountains, eastern France) show a general expansion of population density from the middle Bronze Age to the early Iron Age. This suggests a relative emancipation of protohistoric societies from climatic conditions, probably in relation to the spread of new modes of social and economic organisation.


2500
2000
1500
1000
500 BC
?-2290
2140-2030
vers 1710
1520 -1200/1050
1010
940
830/810 - 350
NIVEAU DES LACS (REGION)
2200-2000
1550-1150
800-400
HABITAT LITTORAL
1801-1510
1120 - 814
1017-965 BC
1079-1038 BC
905-859 BC
CHENS / LEMAN
1801-1774 BC
1645-1570 BC
CONCISE
2790 ± 35 BP
3125 ± 40 BP
2100 ± 100 BP
3330 ± 100 BP
vers 650 BC ?
23795 ± 35 BP
PLUVIS 4
906-814 BC
2700 ± 45 BP
3745 ± 70 BP
2830 ± 55 BP
CHINDRIEUX
?
3475 ± 50 BP
931-825 BC
3450 ± 45 BP
3630 ± 30 BP
TRESSERVE
?
2500
2000
1500
1000
500 BC



931-
825 BC
ca 1550
cal BC
NIVEAU
DU LAC
(m)
GDD5
(°C days)
somme des
températures
quotidiennes
supérieures
à 5°C
(saison de
croissance)
PANN
(mm)
précipitations
annuelles
TANN
(°C)
température
moyenne
annuelle
PSUM
(mm)
précipitations
d'été
MTWA
(°C)
température
moyenne
du mois
le plus
chaud
PWIN
(mm)
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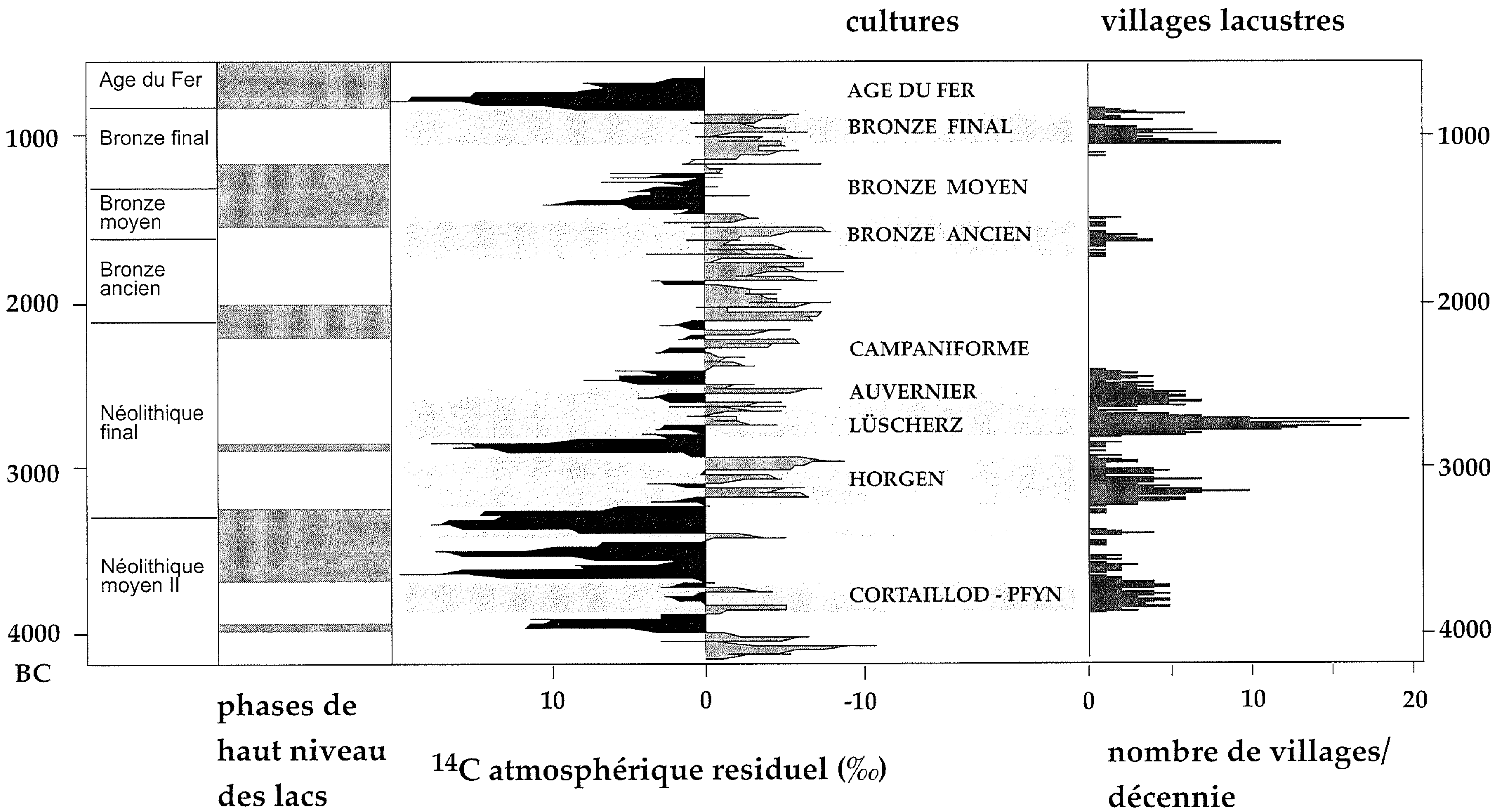
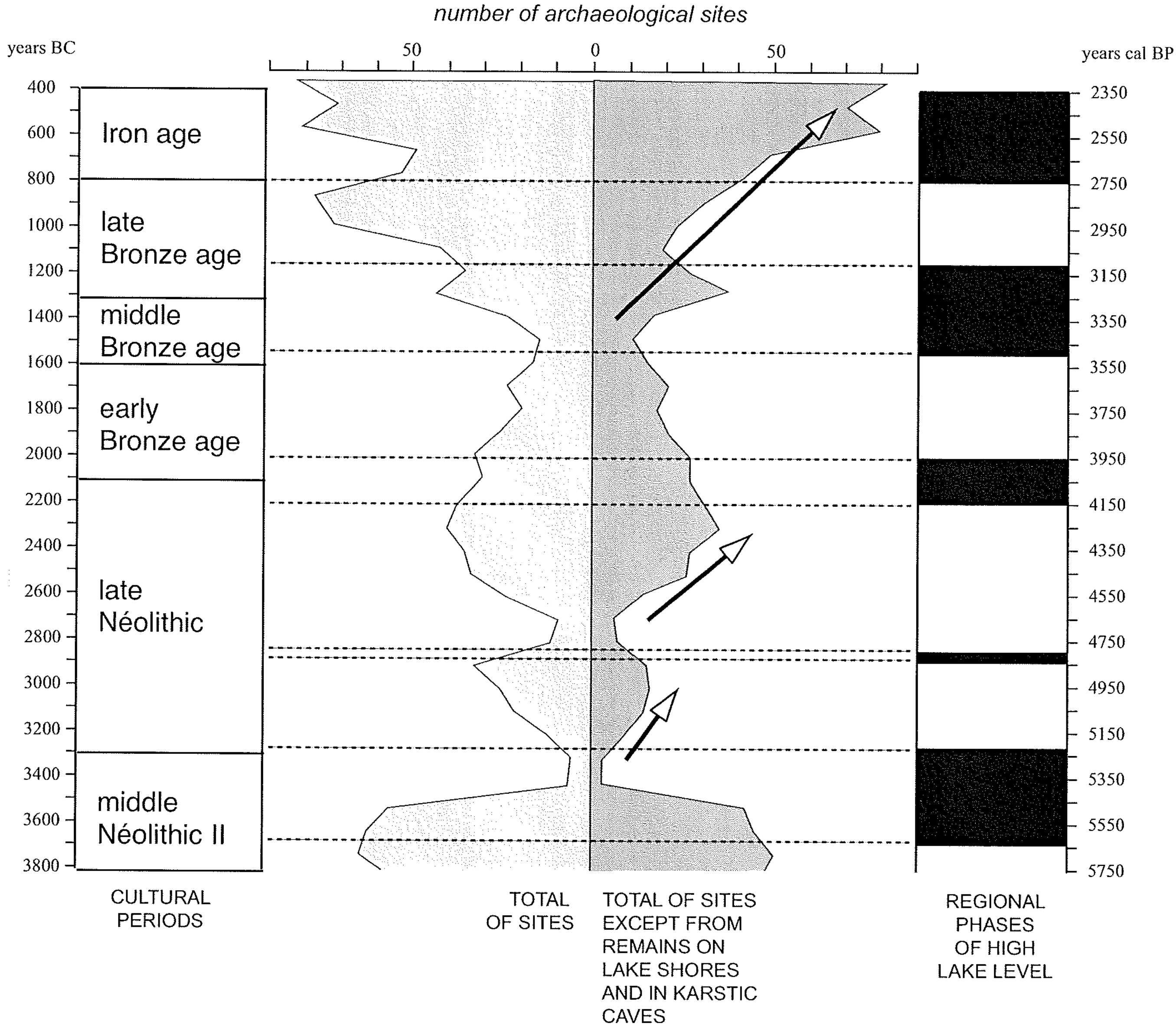
d'hiver
MTCO
(°C)
température
moyenne
du mois
le plus
froid



cultures
villages lacustres
Age du Fer
Bronze final
Bronze moyen
Bronze ancien
Néolithique final
Néolithique moyen II
AGE DU FER
BRONZE FINAL
BRONZE MOYEN
BRONZE ANCIEN
CAMPANIFORME
AUVERNIER
LÜSCHERZ
HORGEN
CORTAILLOD - PFYN
1000
2000
3000
4000
BC
10
0
-10
0
10
20
phases de haut niveau des lacs
14C atmosphérique residuel (‰)
nombre de villages/ décennie



number of archaeological sites
years BC
years cal BP
50
0
50
400
600
800
1000
1200
1400
1600
1800
2000
2200
2400
2600
2800
3000
3200
3400
3600
3800
Iron age
late Bronze age
middle Bronze age
early Bronze age
late Néolithic
middle Néolithic II
2350
2550
2750
2950
3150
3350
3550
3750
3950
4150
4350
4550
4750
4950
5150
5350
5550
5750
CULTURAL PERIODS
TOTAL OF SITES
TOTAL OF SITES EXCEPT FROM REMAINS ON LAKE SHORES AND IN KARSTIC CAVES
REGIONAL PHASES OF HIGH LAKE LEVEL

Friday 21 september

Port-des-Lamberts
Mont-Beuvray

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ORIGINAL ARTICLE

Isabelle Jouffroy-Bapicot · Maria Pulido · Sandrine Baron · Didier Galop · Fabrice Monna · Martin Lavoie · Alain Ploquin · Christophe Petit · Jacques-Louis de Beaulieu · Hervé Richard

Environmental impact of early palaeometallurgy: pollen and geochemical analysis

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Abstract Interdisciplinary research was carried out in mid-level mountain areas in France with the aim of documenting historical mining and smelting activities by means of pollen and geochemical analyses. These investigations were made on cores collected in French peatlands in the Morvan (northern Massif Central), at Mont Lozère (southern Massif Central) and in the Basque Country (Pyrénées). Different periods of mining were recognised from Prehistory to modern times through the presence of anthropogenic lead in peat. Some of these were already known from archaeological dates or historical archives, especially for mediaeval and modern periods. However prehistoric ancient mining activities, as early as the Middle Bronze Age (ca. 1700 B.C.), were also discovered. They had all led to modifications in plant cover, probably related in part to forest clearance necessary to supply energy for mining and smelting.

Keywords France · Palaeometallurgy · Pollution · Lead isotopes · Peatland · Pollen analysis

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Introduction

Recent geochemical analyses have demonstrated that peat bogs are potential archives not only for palaeobotanical purposes, but can also be successfully used to reconstruct historical atmospheric pollution (Shotyk 1996; and special issues of Water, Air and Soil Pollution 100, 1997 and The Science of Total Environment 202, 2002). Recent European studies conducted in the Swiss Jura (Shotyk et al. 1997; Shotyk 2002), Great Britain (Mighall et al. 2002a, b, 2004), Finland (Brännval et al. 1997), Germany (Küster and Rehfuess 1997) and Spain (Martinez Cortizas et al. 2002) suggest that lead can be used to reconstruct atmospheric pollution as it seems to be chronologically retained in hilltop peat and not vulnerable to re-mobilisation, even in non-ombrotrophic peatland (Shotyk 2002). Lead isotopic geochemistry is based on the $^{206}Pb/^{207}Pb$ ratio concomitant with the Pb/Sc or Pb/Al ratio; it indicates the so-called anthropogenic lead (Sc and Al are among elements which behave conservatively in most geological environments). Mining and smelting activities may have affected nearby vegetation through deforestation in response to increasing energy demands. Thus, palynology associated with archaeological knowledge is a powerful tool for a better understanding of prehistoric mining and metal-working (Mighall and Chambers 1993; Küster and Rehfuess 1997; Richard and Eschenlohr 1998).

This paper summarises the results of the first interdisciplinary research in France the aim of which is to document historical mining and smelting activities using pollen and geochemical analyses. These investigations were conducted on peat cores collected in the Morvan mountains (Monna et al. 2004b), in the Lozère mountains in the southern Massif Central (Baron et al. 2005; Ploquin et al. 2003) and in the Basque Country (Pyrénées; Monna et al. 2004a; Fig. 1). They documented palaeoenvironmental impact since the beginning of those activities, in areas where archaeological and historical knowledge indicate mining

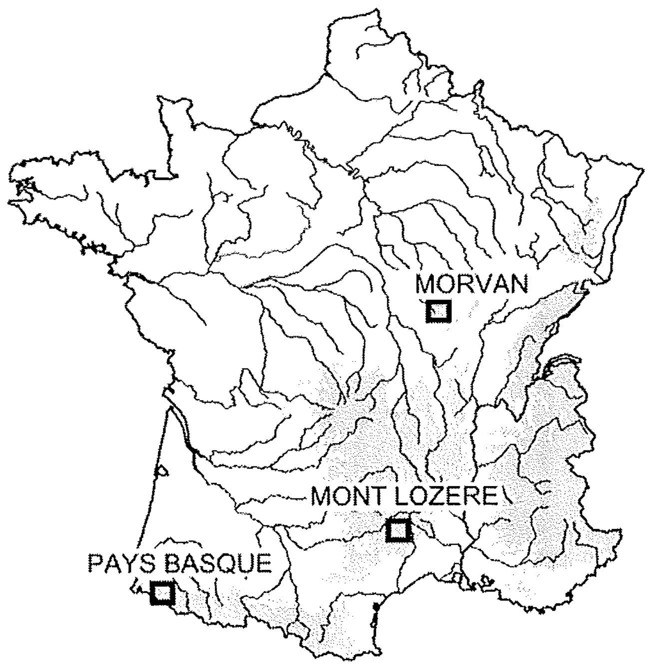

Fig. 1 Location of the sites

and/or smelting activities during several periods and where abundant mineral resources are found.

Sampling sites and their historical context

Mont-Beuvray (Morvan – Burgundy, 820 m a.s.l.) is one of the highest points of the Morvan, northern Massif Central. The Morvan is a Hercynian massif mainly composed of granitic rocks, although volcano-sedimentary terrains (rhyolites and conglomerates) are also exposed. Three main types of mineral deposits were recognised: late Hercynian stratiform barytic and fluoritic outcrops (Lhégu et al. 1982), abundant polymetallic mineralisation (Pb, Zn, Ag) and, to a lesser extent, in conglomerate outcroppings at Mont-Beuvray (Delfour 1978; Marcoux 1986). At Mont-Beuvray, Bibracte, a Celtic hill fort of the Aeduan tribe, became one of the greatest and richest *oppida* in Gaul during the late Iron Age. Geomorphological anomalies such as wide trenches and gullies have recently been discovered and interpreted to be remains of mining excavation. On this basis, archaeologists have assumed that one of the factors that may have attracted early settlers is the abundance of mineral resources.

Mont Lozère (South Massif Central, 1699 m a.s.l.) is located in the Cevennes National Park on the southern edge of the Massif Central (Fig. 1). Mineralisation occurs all around the granitic massif in contact with the Palaeozoic or Mesozoic sedimentary cover: lead is the main metal (galena), while there are also smaller quantities of arsenic, copper, zinc, antimony and silver (Ploquin et al. 2003). On the western side of Mont Lozère, sixty deposits of slag were discovered in an area of 8 km^2, all at an altitude between 1360 and 1430 m above the mineralisation. This slag is metallurgical waste indicating past smelting activities (11th–12th century), mainly from lead and silver extraction (Ploquin et al. 2003). Several slag deposits are located near peat bogs where palaeobotanical and geochemical records are preserved. On the edge of the Narses Mortes peat bog, archaeological excavations have revealed the remains of an old furnace from the mediaeval period.

The High Aldudes Valley (Basque Country) contains abundant mineral resources and was widely mined during Roman times. Archaeological remains dated between the 2nd century B.C. and the 4th century A.D. testify to important metallurgical activity in this area (Galop et al. 2001, 2002; Beyrie et al. 2003). Ores of Fe, Cu, Ag, Sb and to a lesser extent of Pb and Zn, consist of sub-concordant piles or secant veins governed by fractures. Mining activity during the Middle Ages seems to have been very low, while the highest level of production, reaching more than 100 tons of copper, was registered and documented by historical sources before the French Revolution. Exploitation then collapsed during the 18th century A.D. due to a lack of wood supply sufficient for intense exploitation. During the 18th century, this area yielded more than 1200 tons of copper, and around 15 tons of silver (Beyrie et al. 2003). Marginal exploitation is recorded throughout the 19th and 20th centuries.

Material and methods

Sampling

The Port-des-Lamberts peatland (Morvan) is located about 4–5 km from both Mont-Beuvray and the known ore deposit. It covers an area of about 3 ha at an altitude of 700 m a.s.l., is *Sphagnum*-dominated and organically rich at the top. The current surroundings are woodland dominated by beech forest and by a planted spruce grove.

The Narses Mortes peatland (Mont Lozère) is located at an altitude of 1400 m a.s.l. on the western side of Mont Lozère. It is a 21 ha minerotrophic peatland, mainly colonised by peat moss (*Sphagnum*) and characterised by a micro-topography of hummocks (tussocks of *Molina caerulea* or *Polytrichum*) and hollows. Marshy zones with Cyperaceae, *Menyanthes* and *Equisetum*, may develop temporarily. The peatland drains to the south towards the Tarn River. The surrounding vegetation consists of grassland and heath. The only trees around the bog are pines, planted during the second part of the 19th century.

The small peat bog of Quinto Real (Basque Country) is located in the High Aldudes Valley close to the Spanish border at 910 m a.s.l. It covers a surface of 400 m^2 of Paleozoic terrain. The peatland is dominated by *Sphagnum* and Cyperaceae and surrounded by a beech forest and grazing lands.

Pollen analysis

The peat core from Port-des-Lamberts was collected by means of a Russian GIK-type corer, using the conventional two-borehole technique. It consists of about 2 m of organic-rich material. Pollen analysis was performed at a sub-sampling interval of 4 cm. On the Narses Mortes peatland, two perpendicular transects of coring at regular intervals (25 m) permitted description of the geometry of peat accumulation. One sedimentary core was taken using a modified Russian Coring device (Jowsey 1966) at the centre of the peatland (central core, not presented here) where the maximum peat thickness (140 cm) was found. A second core (lateral core) of 138 cm, was collected near the edge of the site, where it was possible to examine the stratigraphy of the sediments through an incision in the peat created by a small, currently inactive stream. This lateral core provided a fine temporal resolution study of the anthropogenic period. The upper part of the core consists of a tussock of *Molinia caerulea*. Pollen analysis was carried out at 2 cm intervals for the upper part (2–74 cm), and 4 cm intervals for the lower part (74–138 cm). In the Quinto Real peatland, a 420 cm core was obtained with a Russian GIK-type corer using the two-borehole technique. Sub-sampling for pollen analysis was carried out at 4 cm intervals in the first metre and at 8 cm intervals in the lower part of the core.

Pollen preparation followed standard procedure, briefly: 10% HCl, 10% KOH, HF and acetolysis. An average of 400 vascular plant pollen grains was counted in each level. Pollen grains were identified with the aid of keys (Faegri and Iversen 1989; Moore et al. 1991), photographs (Reille 1992) and reference to a modern pollen type slide collection. Cyperaceae, spores and aquatic plants were systematically excluded from the pollen sum. In the Port-des-Lamberts peatland core, as at the Quinto Real, *Alnus* was also excluded because its over-representation could have masked the dynamics of other taxa (Janssen 1959; Wiltshire and Edwards 1994).

Radiocarbon dating

Geochemical analysis

The same analytical procedure was applied to the samples from Mont Beuvray and the Basque country. More details about the methodology can be found in Monna et al. (2004a, b). In brief, refractory elements, such as Sc and REE, were measured by Instrument Neutron Activation Analysis (INAA) at Actlabs (Ontario, Canada). For isotopic and heavy metal determinations, about 500 mg of powdered samples were first oxidised with H_2O_2 and digested with a mixture of Suprapur and concentrated HCl, HNO_3 and HF (Merck, Germany) in closed PTX vessels with microwave assistance. An aliquot of the solution was measured by an HP 4500 inductively coupled plasma – mass spectrometer (ICP-MS) to determine Cu, Zn, Cd, and Pb concentrations. Lead from another aliquot was purified using the conventional ionic resin AG1X4 (Biorad) and measured for its isotopic composition by a quadrupole-based HP 4500 (see Monna et al. 1998, 2000 for more details about the chemical procedure and precision). Blanks and reference material standards, including NIST 1547, JSD 1, JSD 2, NIST 1547, PACS-1 and BCSS-1, were also systematically added to each set of unknown samples in order to check accuracy and precision Table 1.

Table 1 AMS radiocarbon dating (*^{14}C-beta counting)

Sites	Depth (cm)	Lab. No.	^{14}C B.P.	Calibrated ages
NM	30	Poz-2011		Modern
NM	41	Poz-7048	850 ± 30	A.D. 1158–1264
NM	50	Poz-2012	1330 ± 35	A.D. 648–773
NM	53	Poz-2014	1265 ± 35	A.D. 667–868
NM	62	Poz-2015	1460 ± 35	A.D. 543–649
NM	71	Poz-2016	1635 ± 35	A.D. 264–5535
NM	90	Poz-1957	1950 ± 40	A.D. 41 B.C.–130
NM	119	Poz-1958	2200 ± 40	382–169 B.C
PL	75	Ly-10942	1070 ± 50	A.D. 888–1028
PL	97	Ly-10943	1460 ± 60	A.D. 441–664
PL	126	Ly-10944	2480 ± 40	790–407 B.C
PL	163	Ly-10945	3117 ± 54	1515–1225 B.C
QR	69–71	Beta-156998	290 ± 40	A.D. 1486–1664
QR	157–159	Ly-10587*	1895 ± 50	A.D. 3—240
QR	229–231.5	Ly-10588*	2645 ± 45	896–787 B.C
QR	283–285	Ly-10589*	3045 ± 70	1485–1051 B.C
QR	357	Beta-156997	4120 ± 40	2876–2501 B.C

Dates were calibrated (2 σ range) using Calib 4.1.3 (PL and QR) or 4.2 (NM) software (Stuiver et al. 1998); NM: Mont Lozere, "Les Narses Mortes" peat core; PL: Mont-Beuvray, "Port-des-Lamberts" peat core; QR: Basque Country, "Quinto Real" peat core; Poz: Poznan Radiocarbon Laboratory (Poland), Ly: Centre des Sciences de la Terre (University of Lyon, France), Beta: Beta Analytic Inc laboratory, Miami

The peat samples from the Mont-Lozère massif were powdered (300 mg) and fused in Pt crucibles along with 900 mg of ultra-pure $LiBO_2$ at 980 °C in an automatic tunnel oven. This methodology is undertaken in routine analyses in the SARM Laboratory at CRPG in Nancy. More details are available in Carignan et al. (2001). For isotopic measurements, dried peat samples (30–300 mg, according to lead concentration in each sample) were dissolved in a Teflon vessel using 2 ml of concentrated HNO_3 and 0.5 ml of 30% H_2O_2. After evaporation at 110 °C, the residue was taken up in 1 ml of concentrated HNO_3, 0.5 ml H_2O_2 and 1 ml of concentrated HF (all Merck Suprapur quality) and left at 80 °C overnight. After the last evaporation, the residue was taken up in 1 ml of 0.9 M HBr. Pb was separated from the other elements by ion exchange using the AG1X8 resin. The lead isotopic composition was measured with a MC-ICP-MS (Isoprobe, micromass, now GV Instruments). The reference materials, NIST 981 Pb and NIST 997 Tl were used to correct any instrumental mass bias. More details concerning the methodology are available in Baron et al. (2005). This technique is based on the relationship measured between Pb and Tl mass bias, according to the empirical technique used by Maréchal et al. (1999) and reported by White et al. (2000) for lead applications. Reference values used for both reference materials were taken from Thirlwall (2002). Repeated measurements of the NIST NBS 981 Pb reference material yielded accurate

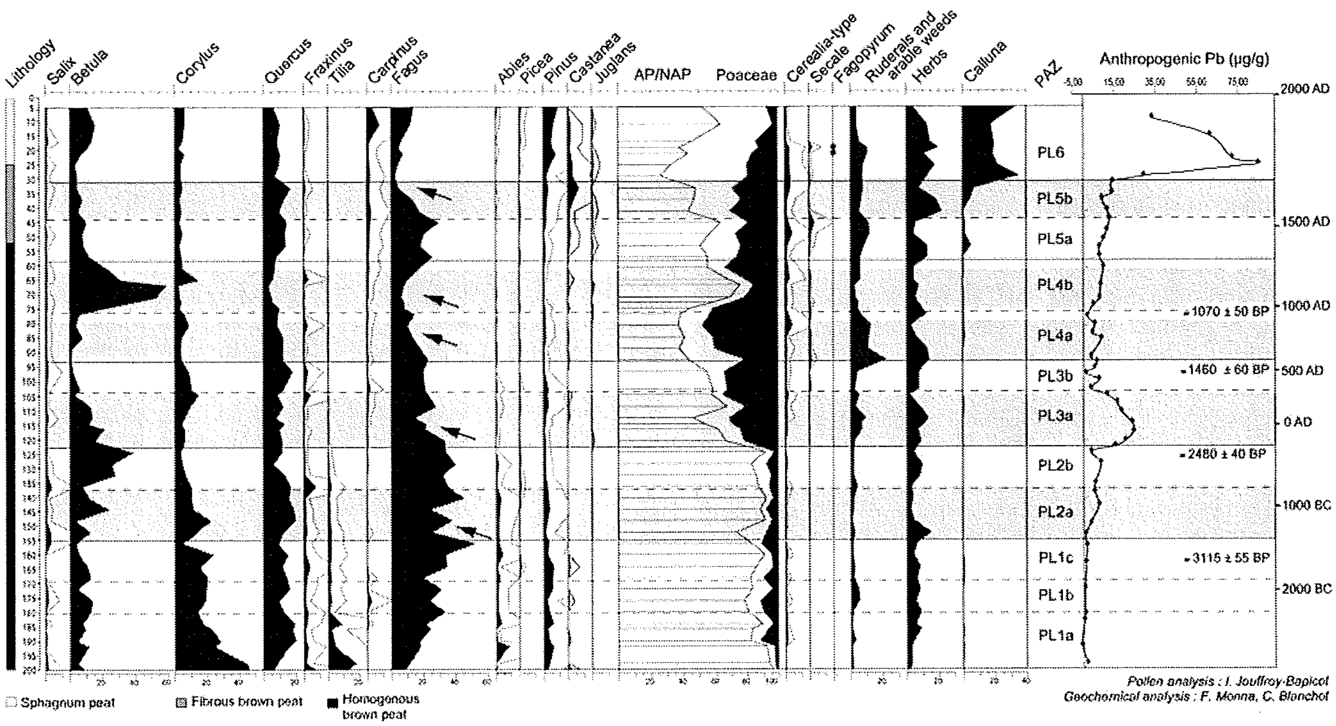

Fig. 2 Simplified pollen diagram (dominant taxa *Alnus* and Cyperaceae were removed) and anthropogenic lead concentration, from the Port-des-Lamberts peatland (Morvan, 700 m a.s.l.)

recalculated values (using the Pb-Tl relationship) with a reproducibility (2 standard deviations) better than 150 ppm for all the reported Pb isotope ratios.

Results

In the Morvan (Fig. 2), according to radiocarbon dates and palynological spectra, the base of the core from Port-des-Lamberts may be estimated to be from the transition Neolithic/Early Bronze Age. This first period (PAZ PL1) is initially dominated by woodland taxa of mesophilous woodland: *Corylus*, *Fagus*, *Quercus* and to a lesser extent *Tilia*. Anthropogenic indicators, such as Cerealia-type and *Plantago lanceolata*, are already present in herbaceous taxa, indicating early human occupation. This was probably for agro-pastoral purposes, because the $^{206}Pb/^{207}Pb$ ratios merely reflected natural mineral matter.

The beginning of PAZ PL2, in the late Bronze Age, shows a drastic drop in woodland taxa, especially *Fagus*. The low percentage of anthropogenic pollen indicators recorded in these levels seems to indicate that the forest clearing was not related to any agro-pastoral extension. It is precisely at this time that the earliest substantial human-derived lead input is noticed. Such a concomitance is an indication of a close connection between metallic contamination and forest clearance. This result suggests that the Mont Beuvray area was, as previously suspected by some archaeologists (Guillaumet 2001), an early mining centre. During the second part of PAZ PL2, the percentage of woodland taxa, dominated by *Fagus*, gradually increases, while anthropogenic lead concentrations remain stable. Human pressure on the forest must have declined at that time – Iron Age, according to radiocarbon dating.

At the beginning of PAZ PL3, *Fagus* collapses again, anthropogenic herbs indicators increase, while anthropogenic lead concentrations peak during the Aeduan civilization (1st century B.C.). This result, together with the numerous metallurgical workshops discovered at Bibracte, may at least partly explain the tribe's wealth. A decline in anthropogenic pollen indicators and lead fluxes marks the beginning of our era. Bibracte was abandoned after the Roman conquest of Gaul, the population leaving the Celtic city to settle in the new city *Augustodunum*, which became the Aeduans' new smelting centre.

Archaeological knowledge from the early Middle Ages is crucially lacking, but anthropogenic indicators are already present and, what is more, their percentages are significant. At the same time, low anthropogenic inputs occurred (PAZ PL4a). Less surprising is the rise in concentration of human-derived lead and the *Fagus* representation, which drops at the beginning of PAZ PL4b, around the 11th century. This event may be synchronous with the great deforestation phase of the Middle Ages, observed on an European scale (Berglund et al. 1996), even if, on a regional scale, there is a lack of archaeological proof and historical information.

Anthropogenic lead inputs continue to increase slightly at the end of PAZ PL5 and reached a maximum at the beginning of PAZ PL6. However, the chronology established from radiocarbon dates does not allow precise dating of this phase. Locally, historical mining during the 18th and 19th centuries is well documented in archives. However the fact that *Fagus* pollen almost totally disappears at that time

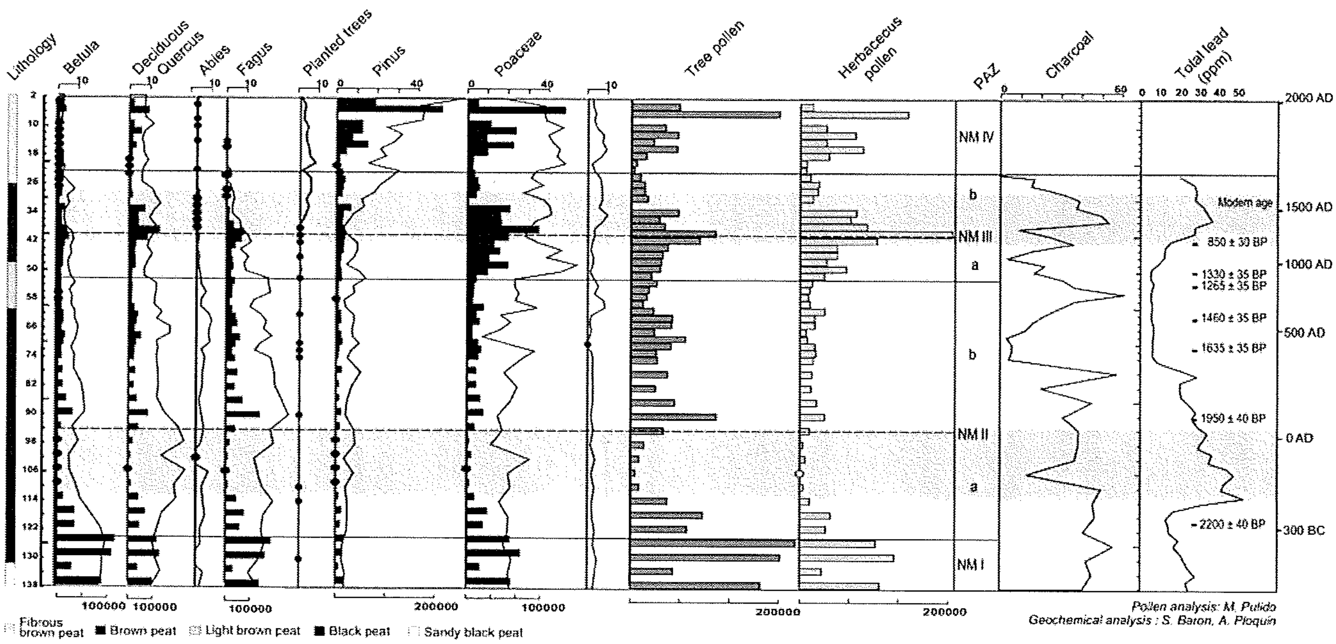

Fig. 3 Pollen diagram from the Narses Mortes peatland (Mont Lozère, 1400 m a.s.l.) and results of geochemical lead analysis. Pollen data are presented in percentages (lines) and concentrations (bars)

may be due to the intense exploitation of the Morvan forests from the 16th to the beginning of the 20th century, in order to supply Paris with firewood. However, this palynological observation may not reflect the reality of the vegetation. Harvesting for firewood was performed by means of coppice selection. There was probably still a beech-dominated forest, but pollen production might have been affected by this activity.

At Mont Lozère (Fig. 3), the age model inferred from the ^{14}C dates from the marginal core from the Narses Mortes peatland (PAZ NMI) suggests that the peat expanded around 2500 years ago. At the bottom of the pollen diagram, a forest environment is deduced from relatively large percentages and the high concentrations of *Betula*, deciduous *Quercus* and *Fagus*. Some grasses (Poaceae) were also abundantly represented in this period, probably due to local contributions from *Molinia*. Former diagrams covering longer periods at Narses Mortes peatland (de Beaulieu 1974) clearly show a moderate opening of the beech forest to the benefit of grasslands and *Calluna* heath, at least from the Early Bronze Age, suggesting a patchy vegetation cover at Mont Lozère. At the same depth in the section, lead concentrations in the peat samples and their corresponding isotopic compositions are close to that of local granite, indicating that no anthropogenic activity occurs.

The beginning of PAZ NM II, Antiquity according to radiocarbon dates, is characterised by a reduction in the abundance of trees. A first major period of deforestation begins at this time. This deforestation was linked to an abrupt increase in lead anomalies, *Fagus* being the taxon most affected. Isotopic compositions for this lead anomaly are the same as those for ores and slag linked to mediaeval metallurgy, but no Iron Age archaeological evidence is present. This zone can be subdivided into two sub-zones (PAZ NM IIa and PAZ NM IIb). In the sub-zone PAZ NM IIa, the tree pollen concentrations decline gradually to reach minimal values at 102 cm. Percentages of *Fagus* and *Betula* decrease equally, whereas those of *Quercus* and *Abies* increase. This synchronous decline in the pollen concentrations of all taxa (trees, shrubs and herbs) undoubtedly results from acceleration in the net peat accumulation rate as compared to the previous period, probably as a secondary effect of deforestation. In sub-zone PAZ NM IIb an increase in pollen concentrations follows. Percentages of *Fagus* and *Betula* increase while those of *Quercus* decrease. At this time, lead concentration decreases probably indicating a reduction in metallurgic pressure. At the end of PAZ NM II, during the early Middle Ages, anthropogenic pollen increases, without any connection to lead concentration and the corresponding lead isotopic composition, thus suggesting agro-pastoral activities.

PAZ NM IIIa is characterised by a second decline in the pollen percentage of *Fagus* and *Betula*. The percentage of *Quercus* remains stable in spite of a sudden reduction in the concentration of this taxon at the end of PAZ NMIII. This second period of deforestation led to the drastic reduction in all tree species (*Fagus*, *Quercus*, *Betula*) found at the end of PAZ NM IIIb, whereas the pollen representation of Poaceae is fairly stable at high percentages and *Pinus* begins to increase at the end of this subzone. PAZ NM IIIa is probably associated with mining activities from the Mediaeval Period (at 42 cm depth). Abundant microscopic charcoal during this period is correlated with the fragmentation of the forest cover in response to anthropogenic fire events, probably linked to local metallurgy The decrease in lead concentrations, recorded in PAZ NM IIIb (at about 34 cm depth), marks the abandonment of metallurgy and an increase in agro-pastoral activities. Indeed, a second maximum in

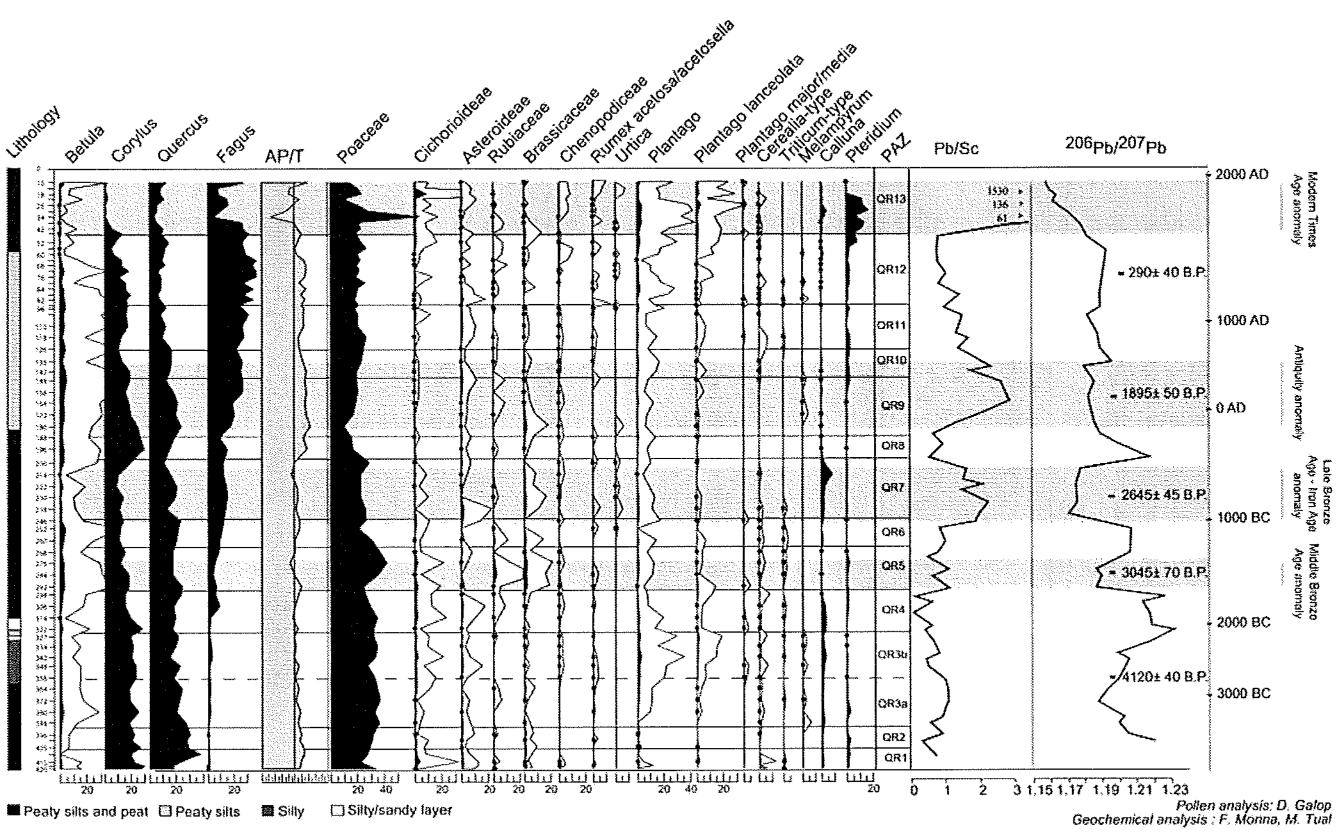

Fig. 4 Pollen diagram from the Quinto Real core (Basque Country, 910 m a.s.l.) with Pd/Sc and $^{206}Pb/^{207}Pb$ ratios

anthropogenic indicators and a continuous curve of planted trees are also recorded. Mediaeval human activities account for the asylvatic nature of the summit of Mont Lozère and for the abnormally low altitudinal tree-limit until the last two decades. At the end of this zone, pine (probably *Pinus sylvestris*) shows a gradual increase. In the uppermost 35 cm, lead isotopic composition indicates that modern industrial inputs overprinted the chemical signals and have thus veiled information from after the 12th or 13th century. The upper 20 cm of the sequence (PAZ NM IV) show pollen spectra marked by a balance between grasses (asylvatic vegetation, probably *Nardus stricta* meadows) and pine (influx from more or less distant plantations since the second half of the 19th century). At the top of the spectra, the pine percentages increase to more than 60% whereas Poaceae drop below 20%, illustrating local invasion by pine.

In the Aldudes Valley (Basque Country; Fig. 4), several phases of metallurgical activities linked with environmental modifications are observed between Late Neolithic and modern times (Middle Bronze Age, Late Bronze Age, Antiquity and finally modern times). In PAZ QR-1 prior to 3000 B.C. during Late Neolithic, the presence of anthropogenic indicators such as Cerealia-type, *Triticum*-type and *Plantago lanceolata* in the pollen record reveals human activities in this area. Later, the decline of oak combined with the slight extension of birch and the slow expansion of herbaceous taxa (mainly Poaceae) registered in PAZ-QR3a suggest a progressive deforestation that might have been the origin of the detrital layer recorded at 310cm depth, around 2000 B.C. At that time the $^{206}Pb/^{207}Pb$ and Pb/Sc ratios indicate natural mineral matter. During the Middle Bronze Age (PAZ-QR5) *Quercus* and *Corylus* declined, whereas a significant increase in herbaceous taxa, such as Poaceae and grazing indicators, is recorded. Isotopic ratios show a close concomitance between forest clearance (AP percentage drops from 60% to below 40%) and the increase in agro-pastoralism (*Rumex* is peaking) and metallic contamination. A similar relationship between lead enrichment and deforestation is noted at ca. 1000–600 B.C. during Late Bronze Age/Iron Age transition (PAZ-QR7). During this period the pollen record indicates a removal of oak forest while traces of agro-pastoral activities (*Plantago lanceolata, Plantago major/media, Rumex*, Cerealia-type) decrease or are absent. In this case, deforestation does not parallel agro-pastoral extension and may be interpreted as the result of increasing energy demands for mining and smelting purposes.

Another major anthropogenic phase is pinpointed by Pb/Sc and $^{206}Pb/^{207}Pb$ ratios from ca. 200 B.C. to A.D. 200 (PAZ-QR9 and QR 10). During this period, *Quercus* and *Corylus* decrease while *Fagus* seems to spread. Moderate signs of deforestation appear again without any pollen indication of significant agricultural extension. The Pb/Sc peak observed in the core corresponds well with the exploitation of iron, copper, silver and lead from the metallurgical and mining sites of the Baïgorry Valley, well known in Antiquity (Beyrie et al. 2003). During this period the decline of oak can be explained by deforestation for metallurgical operations. Moreover, anthracological analysis performed in shaft-furnaces located near the peatland established that charcoal production within the valley focused on this species (Galop et al. 2002). After a long decline throughout the Middle Ages (PAZ-QR11 and QR12), Pb/Sc ratios increase markedly from late 16th century and early 17th century A.D., coinciding with the decrease in $^{206}Pb/^{207}Pb$ ratios (PAZ-QR 13). This period is an intense phase of metallurgical activities in the Basque Country, particularly in the Baïgorri Valley where the copper foundry started operating in 1747. Forest taxa (*Fagus*, *Quercus*) declined as metalworking peaked, indicating intense forest clearance linked with wood charcoal production and pastoral activities suggested by the increase of anthropogenic indicators. In this area, forests were dedicated to charcoal production, as demonstrated by abundant charcoal-kiln remains in current forest and pasture areas of the Aldudes Valley.

Discussion

In the Morvan, as in the Basque country and the southern Massif Central, different phases of palaeometallurgy are recognised according to the presence of anthropogenic Pb in peat. Those activities of extraction and/or smelting of different metals (copper, silver or gold) would have emitted enough lead-enriched dust and gases into the atmosphere to be retained in surrounding environments.

The different phases of palaeometallurgy, recognised from the presence of anthropogenic Pb in peat, induced major modifications in plant cover. This is probably related in part to the forest clearance necessary to supply energy for mining and smelting. Vegetation cover may have been drastically affected by selective deforestation, affecting especially *Fagus* (beech) in the Morvan (Monna et al. 2004b) and at Mont Lozère (Lavoie et al. in press), while *Quercus* (oak) was preferentially used for charcoal production in the High Aldudes Valley (Galop et al. 2001; Fig. 2). Beech is well known as a good fuel for energy production.

According to the radiocarbon dates, some periods of high metallurgical activities were related to archaeological evidence of mining or smelting activities. However, some evidence of palaeometallurgy, marked by pollution and a significant fall in the pollen percentage of arboreal taxa, could not be related to archaeological knowledge, thus giving new data concerning human settlement in these areas. This might explain the forest clearance with no increase in pollen anthropogenic indicators linked to agro-pastoral activities. The results of these paleoenvironmental studies provide evidence of proximal metal activities for periods in which archaeological knowledge is lacking, for example the Bronze Age in both the Morvan and the Basque Country and during Antiquity at Mont Lozère. Where and when archaeological data or archives do exist, these paleoenvironmental studies also provide a reliable chronology and a new set of data for mining or smelting activities. The Morvan and Aeduan metallurgy, Mont Lozère during the Middle Ages and the Basque Country during both Antiquity and modern times, are all examples.

The studies demonstrate the strength of the environmental impact on soil pollution and vegetation of pre-industrial activities, in regions which are nowadays among the less industrialised areas of France. Mining and smelting activities have had strong and long lasting effects on the forest cover. In the case of soil pollution at Mont-Beuvray, for example, about 20% of the total anthropogenic lead found there was deposited before our era, and probably about 50% of it before the 18th century (Monna et al. 2004b). The Pb concentrations in workshop-soils at Mont-Lozère, are of the same order as modern ones, but spread over a smaller area. This heritage should be taken into account in the study of present and future environmental problems.

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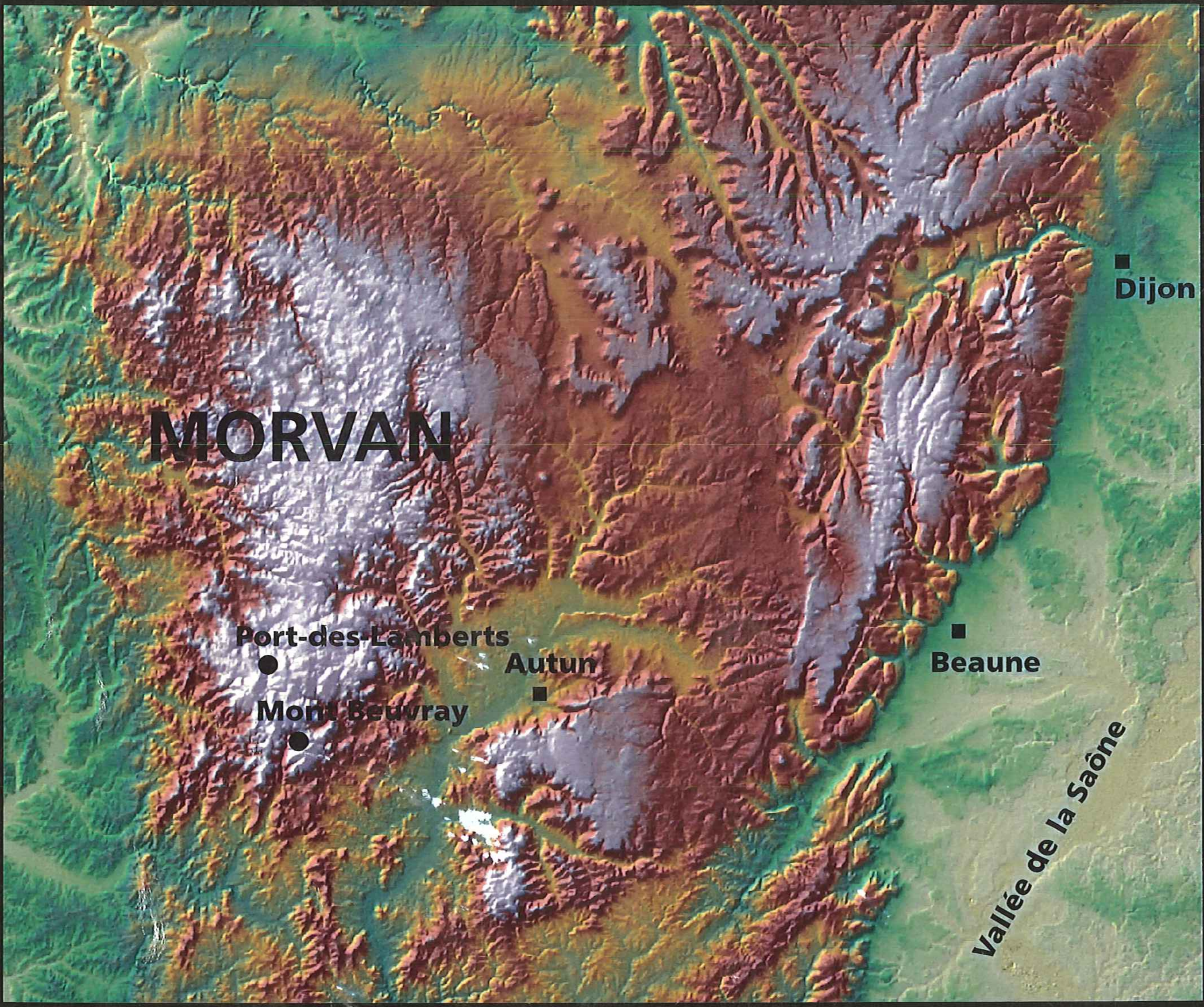
Notes



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