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Non-reversible geosystem destabilisation at 4200 cal. BP: Sedimentological, geochemical and botanical markers of soil erosion recorded in a Mediterranean alpine lake

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Abstract

A 144-cm-long core was obtained in Lake Petit (2200 m a.s.l., Mediterranean French Alps) in order to reconstruct past interactions between humans, the environment and the climate over the last five millennia using a multidisciplinary approach involving sedimentological, geochemical and botanical analyses. We show a complex pattern of environmental transformation. From 4800 to 4200 cal. BP, podzol-type soils progressively developed under forest cover. This stable situation was interrupted by a major detrital pulse at 4200 cal. BP that we consider as a tipping point in the environmental history. At this point, pedogenetic processes drastically regressed, leading to the development of moderately weathered soils. More frequent detrital inputs are recorded since 3000 cal. BP (AD 1050) as the human impact significantly increased in the catchment area. We conclude that destabilisation of the environment was triggered by climate and exacerbated by human activities to a stage beyond resilience.

Keywords

4.2 ka event, human pressure, Mediterranean Alps, soil erosion, tipping point, XRF core scanner

Introduction

Soil erosion and degradation are often regarded as major characteristics of the Mediterranean area (García-Ruiz, 2010). Whereas questions remain open on the historical causes of modern generalised soil disequilibrium that lead to colluviation processes, a debate has begun regarding the respective importance of climatic and human factors in erosion processes (Roberts et al., 2011). Indeed, human practices and meteorological events are well-known factors forcing long-term soil erosivity and soil erodibility (Hatfield and Maher, 2009; Nearing et al., 2005; Strunk, 2003). However, deciphering these forcing factors over the Holocene period to provide new insight on landscape sensitivity and threshold effects remains a challenge (Butzer, 2005), especially in high-altitude areas where the climatic signal is amplified as a result of large topographical gradients (Beniston, 2003) and traditional land-use activities such as domestic livestock grazing, forest clearance and cultivation of marginal land (Röpke et al., 2011; Walsh et al., 2007). Palaeoenvironmental reconstructions can throw light on this debate provided they rely on accurate proxies and reliable chronologies.

Lake sediments are well-recognised archives that provide long-term perspectives in the deciphering of geosystem trajectories (Magny et al., 2013) by recording a broad range of proxies responding more or less directly to changes in environmental conditions (Birks and Birks, 2006). One key advantage of this multiproxy approach is to avoid age–depth uncertainties when various environmental proxies are investigated on the same cored profile and at a high temporal resolution (Fritz, 2008). This enables the determination of a common time frame of geosystem changes, including those related to weathering and soil development as a measure of stability or instability (synchronism, time-gap, tipping point).

The geophysical and geochemical properties of lacustrine bulk sediments have been shown to be valuable in tracking erosion processes (Enters et al., 2008; Giguet-Covex et al., 2011;

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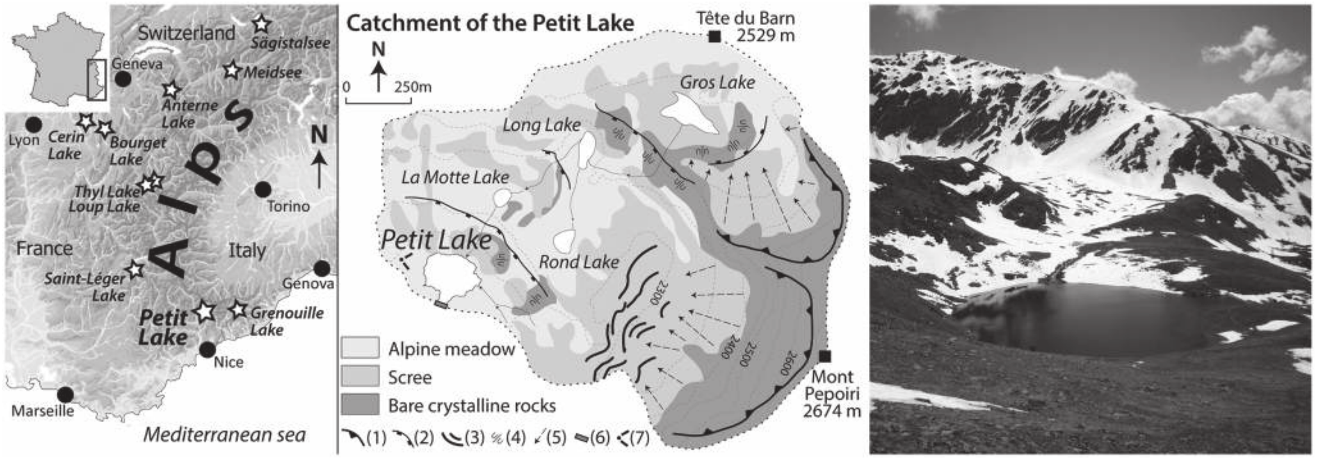

Figure 1. Location map of the study site and other lake records referred to in the text, main geomorphological characteristics of the catchment and photography of Lake Petit. (1) glacial cirque, (2) glacial step, (3) moraine, (4) polished bedrock, (5) active debris slope, (6) small dam built in 1947 and (7) photography viewpoint.

Ohlendorf et al., 2003) and changes in hydrolysis regimes involved in soil development (Arnaud et al., 2012; Koinig et al., 2003). Warm climatic periods potentially intensify chemical weathering (White et al., 1999), whereas forest vegetation is deemed to reduce soil erosivity (Kauppila and Salonen, 1997). The temporal variation of these weathering processes has been investigated through analysis of major and trace elements (e.g. Mourier et al., 2010). However, element analysis is extremely time-consuming, and this is an obstacle to systematic whole-core analysis. High-resolution measurement techniques have been developed in parallel with much less time-consuming devices. One routinely used technique is that of sediment magnetic properties, which are in many cases characteristic of the detrital fraction (DF) and may reflect changes in particle size (Dearing et al., 2001; Oldfield et al., 2003). However, other sediment compounds and their elemental composition cannot be accessed using this technique. X-ray fluorescence (XRF) core scanning (Wilhelm et al., 2012) and Fourier-transform infrared spectroscopy (FTIRS; Vogel et al., 2008) offer fast tools to quantify at very high resolution the elemental geochemistry (lithogenic major elements) and bulk composition (organic matter, biogenic silica), respectively.

For the first time in the southern French Alps, we used these techniques as part of a high-resolution multiproxy investigation of a core from an alpine Mediterranean lake (Figure 1) covering the last 5000 years. The altitudinal position of Lake Petit above the current treeline and the presence of archaeological remains of both pastoral and mining activities in the catchment offer exceptional conditions for the investigation of past interactions between environment, climate and humans using the lake's sediment archive. Major elements, biogenic silica and organic matter from the core were analysed and compared with pollen and botanical data in order to decipher changes in weathering and erosional regime in relation to climate fluctuations, slope vegetation dynamics and human impact on the fragile mountain environment.

Study site

Lake Petit (2200 m a.s.l.) is a small circular lake (diameter: 150 m and depth: 7 m) in the southwestern extremity of the French Alps in the Mercantour-Argentera Massif (N 44°06.789; E 7°11.342; Figure 1). This glacier-inherited lake is surrounded by mountains reaching 2670 m, and is part of the Millefonts catchment. This region of the Alps is less than 40 km from the Mediterranean Sea and is influenced by both Mediterranean and continental alpine climates. Mean annual precipitation in the Mercantour Massif at 1800 m a.s.l. is about 1340 mm and mostly occurs as rainfall in spring and autumn. Snow covers the catchment *c*. 138 days per year. Mean temperatures range from 0.4°C in winter to 9.9°C in summer. The glacier-inherited catchment is characterised by gentle south-facing slopes (*c*. 20°) contrasting with north-facing vertical rock bars. The basement consists of Palaeozoic igneous rocks (gneiss and migmatite) mainly covered by scree. Lake Petit is the largest of four lakes seasonally connected by streams running in scree and moraines. Lake Petit is connected to two water-seepage zones supplied by snowmelt during summer. In this part of the Alps, the Mediterranean and supra-Mediterranean vegetation belts reach their highest altitudes (Ozenda, 1985). Within the study area, the treeline reaches 2100 m and is characterised by a sparse presence of larch. The slopes are covered by *c*. 40-cm-thick soils mostly developed on moraine deposits. We distinguish two types of soils in the Millefonts catchment. Soils developed under an alpine dry acidic meadow are characterised by a gradual transition between three silty-sand dry horizons that become richer in stones towards the deeper horizons (Cambisol (drystric) sensu IUSS Working Group WRB, 2006). Soils developed under dwarf shrub formations of Ericaceae are characterised by well-contrasted horizons from dark brown lumpy surface horizons to ochre silty-sand deeper horizons (Entic Podzol sensu IUSS Working Groups WRB, 2006). At lower altitudes, the mountain vegetation belt shows an alpine to Mediterranean combination of species and is characterised mainly by *Pinus cembra, Abies* and *Picea* in association with *Olea* and *Buxus*.

Materials and methods

A 144-cm-long core (PET09P2) was retrieved from the deepest part of Lake Petit in 2009 using an UWITEC gravity coring system. The sediment core PET09P2 is composed of homogenous yellow to greenish diatomite with millimetre-thin brownish diatomite-clay diffuse laminations. After core opening and lithological description, we continuously cut one of the core halves into 144 volumetric sediment samples to measure dry density (g/cm^3) of the bulk sediment and to concentrate fossil sporo-pollinic content. The other half core was scanned with the XRF method in order to conduct continuous semi-quantitative measurements of major elements, further calibrated by quantitative concentration analysis.

Dating

^{210}Pb and ^{137}Cs short-lived radionuclide gamma decay was analysed at a 1-cm interval over the uppermost 20 cm (*Laboratoire de*

Table 1. Radiocarbon ages of core PET09P2. The maximum probability ^{14}C age is typed in bold between the 2σ interval.

Depth (cm)	Laboratory code	^{14}C age BP	2σ calibrated age (cal. yr BP)	2σ calibrated age (cal. yr AD/BC)	Material
26	Poz-32578	1240 ± 40	1068 (**1175**) 1270	AD 680 (ad **775**) AD 882	Terrestrial debris
39	Poz-39213	1720 ± 30	1554 (**1680**) 1704	AD 246 (ad **270**) AD 396	Wood – *Larix*
58	Poz-35509	1890 ± 30	1733 (**1825**)1894	AD 56 (ad **125**) AD 217	Wood – conifer
94	Poz-32576	3620 ± 40	3835 (**3910**) 4080	2131 BC (**1960** bc) 1886 BC	Terrestrial debris
111	Poz-35507	3855 ± 35	4155 (**4250**) 4411	2462 BC (**2300** bc) 2206 BC	Scale – conifer
135	Poz-39212	4125 ± 35	4528 (**4610**) 4820	2871 BC (**2660** bc) 2579 BC	Twig – Ericaceae
137	Poz-32577	4110 ± 35	4455 (**4610**) 4818	2869 BC (**2660** bc) 2506 BC	Twig fragments

Glaciologie et Géophysique de l'Environnement (LGGE), France). Seven ^{14}C ages were obtained from terrestrial macro-remains throughout the core (Table 1), and dating was carried out by the Poznan Radiocarbon Laboratory. ^{14}C ages BP were calibrated using the IntCal09 calibration curve (Reimer et al., 2011). An age–depth model was computed using a monotonic smooth spline (type 4, smooth 0.3) on the clam module (Blaauw, 2010) taking into account the probability density function of the ^{14}C ages.

Grain-size, organic matter, biogenic silica and major elements

Grain-size distributions were measured by laser diffraction using a Malvern Mastersizer S at 5 mm intervals throughout the core. Samples were ignited at 550°C prior to grain-size analysis to remove organic matter. Distributions were standardised, and the deviation of the grain-size class [1.95, 2.28] µm was calculated to investigate abundance variation of clays. Total organic carbon (wt%TOC) was analysed on 24 samples by Rock-Eval Pyrolysis using a 'Turbo' Model RE6 pyrolyser (*Institut des Sciences de la Terre, Université d'Orléans* (ISTO), France). Quantification of biogenic silica was carried out on 28 discrete samples following wet-alkaline leaching according to DeMaster (1981). Dissolved silica concentration was measured using a molybdate-yellow spectrophotometry with a Jasco V-650 Spectrophotometer. Three replicates were measured on five samples giving a mean standard error of 4.9%. Concentration of biogenic silica is expressed as $\%SiO_{2\,biog}$ ($\%SiO_{2\,biog} = 2.139 * \%Si_{opal}$ according to Mortlock and Froelich, 1989). Bulk sediment concentrations in SiO_2, Al_2O_3, Fe_2O_3, MnO, MgO, CaO, Na_2O, K_2O, TiO_2 and P_2O_5 were measured on 20 samples by mass spectrometry using an inductively coupled plasma–atomic emission spectroscopy (ICP-AES) Thermo X7 after $LiBO_2$ alkaline fusion (Carignan et al., 2001; Murray et al., 2000) by the *Service d'Analyse des Roches et des Minéraux* (SARM) laboratory. Prior to the analyses, the aliquots were heated at 1000°C during 7 h, giving the loss-on-ignition weight. Abundances of major elements are expressed as oxide weight percentages (wt%). The DF of SiO_2 ($SiO_{2\,detr}$) is calculated assuming a constant ratio between aluminium and silica detrital phases in the upper continental crust ($(Si/Al)_{UCC} = 3.83$; McLennan, 2001) as:

$$SiO_{2\,detr}\left(\text{wt }\%\right) = \%Al_2O_{3\,\text{Sediment}} * \left(Si / Al\right)_{UCC}$$

The total DF (%DF) is calculated as the sum:

$$\begin{aligned} DF\left(\text{wt }\%\right) = {} & \%SiO_{2\,detr} + \%Al_2O_3 + \%Fe_2O_3 + \\ & \%MnO + \%MgO + \%CaO + \%Na_2O \\ & + \%K_2O + \%TiO_2 + \%P_2O_5 \end{aligned}$$

As the dilution effect between biogenic silica and the detrital silicate phase is predominant, we investigated the chemical composition of the detrital silicate phase. The $X_{Silicate}$ (mol/100 g) is the abundance of the element X in the DF:

$$X_{Silicate} = X / DF\ (\%)$$

To investigate the chemical composition of the detrital phase, we used the Chemical Index of Alteration (CIA; Nesbitt and Young, 1982):

$$\begin{aligned} CIA\ (\%) = {} & \%Al_2O_3 / \\ & \left(\%Al_2O_3 + \%CaO + \%K_2O + \%Na_2O\right) \end{aligned}$$

High-resolution analysis

To achieve high-resolution analysis, within-core calibrations were constructed using quantitative discrete samples. We used the FTIRS method (Vogel et al., 2008) to predict $SiO_{2\,biog}$ and TOC. Measurements were carried on a Nicolet 380 Smart Diffuse Reflectance spectrometer on 144 samples providing continuous resampling at cm-resolution. XRF core scanning was carried out using an ITRAX Core scanner. Relative element abundances of K, Ti, Fe, Rb, Si and Ca were measured using a molybdenum source at an increment of 2 mm over an integration time of 60 s (40 kV and 35 mA).

Pollen analysis

Pollen chemical extraction was conducted following the protocol of Bennett (1990) on 117 volumetric samples. Taxonomic identification was performed under a 500× magnification oil immersion microscope using identification keys (Moore et al., 1991) and photographs (Reille, 1995). A minimum of 300 pollen grains were counted. Taxa percentages were calculated on the total pollen sum. We present selected pollen taxa that are characteristic of local pollen rain. Pollen of *Rumex, Urtica, Mentha, Plantago lagopus, Plantago lanceolata, Plantago coronopus, Plantago major* and the Chenopodiaceae are summed to synthesise the anthropogenic-related taxa. The abundance of *Botrychium* spores is expressed in sediment concentration. Microfossils of *Pinus* tracheid and conifer stomata were also considered. The pollen diagram was plotted using the software C2 (Juggins, 2007).

Results and interpretation

Chronology

The ^{210}Pb and ^{137}Cs profiles show a common pattern (Figure 2a). The top 5 cm are enriched in radionuclides, and the values drop to nil concentrations deeper down the core. Possible vertical sediment mixing (e.g. bioturbation) could explain the flattening in the top 5 cm of radionuclide activity. Assuming a constant initial ^{210}Pb concentration and constant accumulation rate (CFCR model; Robbins, 1978), the mean sedimentation rate is estimated at 0.7 mm/yr. This value is in agreement with the beginning of nuclear weapons testing and the associated ^{137}Cs pollution (AD 1952 ± 2; Longmore, 1982) at 5 cm depth. Disappearance of laminations in the top 5 cm may be the sedimentological expression of the small dam

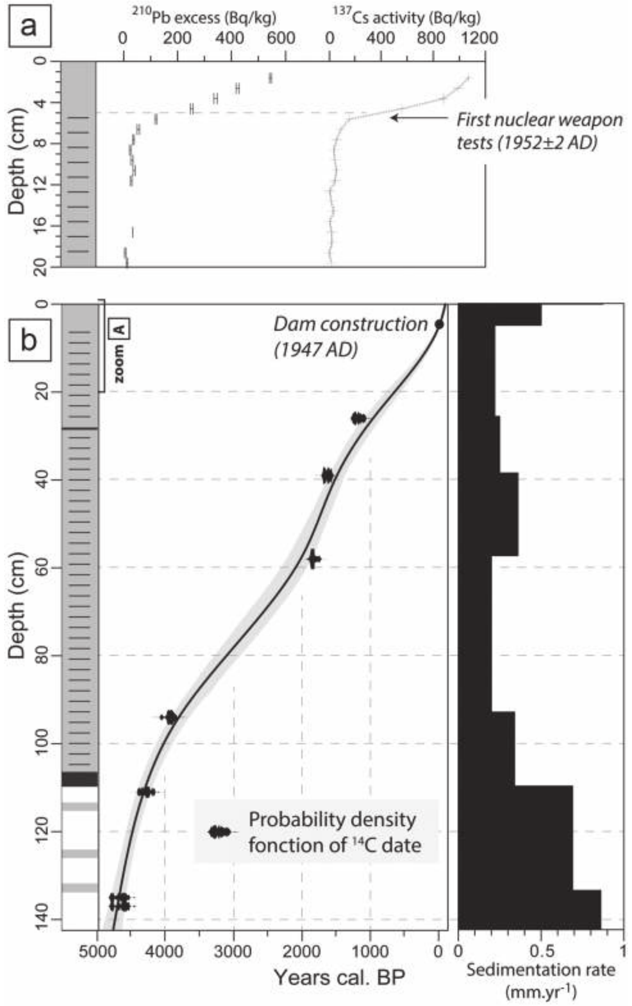

Figure 2. Age–depth relationship and sedimentation rate of core PET09P2. (a) ^{210}Pb and ^{137}Cs levels in the uppermost 20 cm and (b) age–depth model (black line) and the 95% confidence interval (grey envelope).

constructed at Lake Petit in AD 1947 (Beniaminio, 2006). The ^{210}Pb and ^{137}Cs data are consistent with such an interpretation. This may signify that the sedimentation rate changed at 5 cm, which prevents extrapolation of the ^{210}Pb sedimentation rate deeper. Hence, the PET09P2 age–depth model (Figure 2b) has been constrained at 5 cm up to the year AD 1947 and by the seven ^{14}C ages in Table 1. The Lake Petit record covers the last 4700 years. The sedimentation rate is relatively low between 5 and 105 cm (mean 0.24 mm/yr). Higher sedimentation rates are recorded from 0 to 5 cm (0.5 mm/yr) and from 105 to 144 cm (0.64 mm/yr).

Calibration models of TOC, $SiO_{2\ biog}$, DF and CIA

Two partial least squares (PLS) models were built in order to predict %$SiO_{2\ biog}$ and %TOC using the FTIR spectra at a resolution of 1 cm. The within-core FTIRS-inferred (−$FTIRS_{inf}$) models (see Supplementary Materials 1, available online, for details on these calibration models) have a significant quality of adjustment (%$SiO_{2\ biog}$ − $FTIRS_{inf}$: $r^2 = 0.80$, standard deviation error of residuals (SDE_{res}) = 5.99%, $n = 28$, %TOC-$FTIRS_{inf}$: $r^2 = 0.94$, $SDE_{res} = 0.62\%$, $n = 24$). Although the results of opal concentration between FTIRS and conventional leaching technique present a reasonable agreement, the estimates based on FTIRS exhibit lower intensity in variations. The observed offset could be related to the fact that some of the refractory forms of amorphous silica could be dissolved by leaching (Saccone et al., 2007), whereas biogenic silica could be confounded with other compounds absorbing in the same infrared region (Rosen et al., 2010).

The XRF element profiles were compared to the results obtained by quantitative measurements (ICP-AES) of discrete samples ($n = 20$) collected within the core. Large variations in dry density were observed, ranging from 0.1 to 0.6 g/cm^3, with low values in diatom-rich units and high values in more clayey sediment. Changes in dry density have been pointed out as inducing important scatter bias in cross-plots of XRF core scanner outputs and element concentrations. Weltje and Tjallingii (2008) propose a calibration model based on the log-function of the XRF output intensities, which reduces the sensitivity to the dilution effect. XRF intensities of lithogenic elements K, Ti, Fe and Rb are highly cross-correlated (minimum cross-correlation = 0.92). Since Ti shows a good signal-to-noise ratio, it was used as a DF proxy. Titanium dioxide (TiO_2) and %DF are highly related ($r^2 = 0.99$). Ln(Ti/Si) is correlated to the %DF ($n = 20$, $r^2 = 0.89$, $SDE_{res} = 5.4\%$; see Supplementary Materials 2, available online, for details on these calibration models). Since variations in calcium content provide the clearest indications on the CIA index (alone, CaO explains 81% of the CIA variance), we hypothesise that the Ln(Ti/Ca) ratio could summarise the CIA index information. The relationship between the Ln(Ti/Ca) XRF intensities and the CIA is linear and positive, and both are highly related ($n = 20$, $r = 0.90$).

Bulk sediment composition

Core PET09P2 (Figure 3a) is organic-rich (9% mean of TOC) and has a high abundance of mainly biogenic silica (mean 65% of $SiO_{2\ biog}$). Important differences in dry density are observed between the diatom-rich sediment (0.15 g/cm^3) and the most clayey unit (0.7 g/cm^3). All grain-size samples show bimodal distributions mainly characterised by well-preserved diatoms at 25 µm (*Pseudostaurosira robusta* and *Staurosirella pinnata* are the most common species) and diatom fragments (~0.4 µm).

The cross-correlations (Table 2) between Al_2O_3, Fe_2O_3, MnO, MgO, Na_2O, K_2O, TiO_2 and P_2O_5 are highly positive and represent the lithogenic DF (%DF). Lithogenic elements are negatively correlated with $SiO_{2\ biog}$ and to lesser extent with TOC content. The TOC concentration is partly negatively related to lithogenic compounds (Table 2). TOC is a mixture of terrestrial and lacustrine organic debris (presence of diatoms, green algae, chironomid head capsules, wood fragments, spores, pollen and amorphous soil particles). CaO is not correlated with the other lithogenic elements (Table 2). As shown by the absence of absorbance in the carbonate-related bands in infrared spectroscopy, we postulate that the sediment is carbonate-free in agreement with the igneous catchment geology. Hence, CaO concentrations are related to the lithogenic fraction but present a different pattern compared to other lithogenic elements. The concentration of %DF increases from the bottom to the top of the core (20–50%, respectively; Figure 3a). At 104–112 cm, %DF shows peaks of up to 90%. In contrast to %DF, %$SiO_{2\ biog}$ decreases from *c.* 70% at depths of 120 cm to *c.* 30% at this depth. The geochemical composition of sediment is primary controlled by dilution effects between the TOC, the biogenic silica and the clay minerogenic inputs.

Element abundance within the DF

Figure 3b depicts element concentration within the total DF. The elements are grouped into two main negatively correlated ($r = -0.9$) compositional phases: [Al, Ti, Mg, Na, K] and [Ca, Fe, Mn, P]. Al is strongly correlated to Ti ($r = 0.94$), a relatively inert lithogenic element (Mackereth, 1966), and represents the

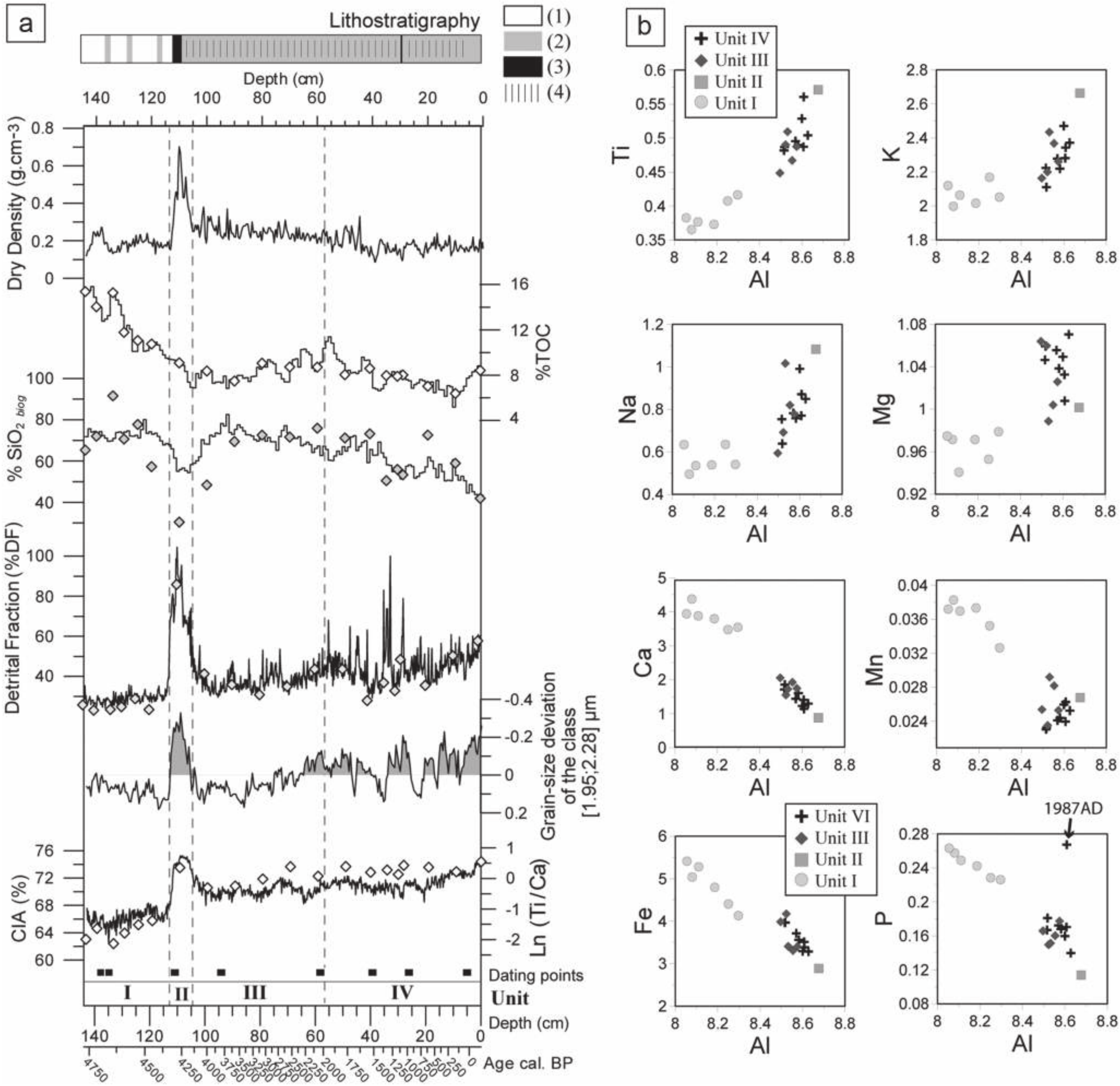

Figure 3. (a) Lithostratigraphy and vertical profile of bulk sediment: (1) pure diatomite, (2) diatomite-clay, (3) clay-diatomite and (4) diffuse laminations. Diamond dots correspond to quantitative values used to model calibration (line) for TOC, $SiO_{2\ biog}$ and detrital fraction. The CIA is plotted against the raw data Ln(Ti/Ca). (b) Harker's diagram of major element concentration (mol/100 g) versus aluminium (mol/100 g) within the terrigenous fraction.
CIA: chemical index of alteration; TOC: total organic carbon.

alumino-silicate phase. The transition between Units I and II (Figure 3b) is characterised by a strong decline in concentrations of [Ca, Fe, Mn, P] that contrasts with the increase of the alumino-silicate phase (Figure 3b). At lower magnitudes, Mg, K, Na and Ti accumulated in the DF of the sediment compared to Al in Unit I. The CIA of Unit I (64%) is similar to the signature of poorly altered igneous rocks rich in primary minerals (Nesbitt and Young, 1982). From Unit I to Unit II, CIA increases from 64% to 72% (Figure 3a). This transition is not associated with modification of grain-size, and only the abundance of occupied volume by class varies (especially in the class 1.95–2.28 µm). The lithogenic fraction in Units II, III and IV has a signature of poorly altered secondary minerals (mean CIA of 72%; Nesbitt and Young, 1982). A noteworthy maximum value of CIA (75%) in Unit II indicates the highest alteration degree of the DF throughout the sequence.

Vegetation changes

The pollen record (Figure 4) is dominated by the *Pinus sylvestris*-type with relative frequencies varying from 30% to 60% of the total pollen count. Pollen of *Pinus* are often over-represented in temperate mountain environments because of its very large wind-dissemination and massive production (Coûteaux, 1982). As a result, the *Pinus* curve must be interpreted with caution since its over-representation may be accentuated in open grassland environments. Nevertheless, terrestrial macro- and microfossils of trees and shrubs (wood, twigs, barks, leaves and needles, *Pinus* tracheid and conifer stomata) found in the sediment core are good indicators of the local presence of woodlands or at least of isolated trees such as larch and pine (*P. sylvestris* and/or *Pinus uncinata* and *P. cembra*) between 4800 and 1500 cal. BP.

The Local Pollen Assemblage Zone I (LPAZ 1) is dominated by arboreal pollen (60%). Conifer trees were present in the lake

Table 2. Correlation matrix of concentration in major elements (ICP-AES measurements) and TOC. 1% significant correlation coefficients are shown in bold, n = 20.

	Al_2O_3	Fe_2O_3	MnO	MgO	CaO	Na_2O	K_2O	TiO_2	P_2O_5	$SiO_{2\ biog}$
Fe_2O_3	**0.94**	1.00								
MnO	**0.92**	**0.93**	1.00							
MgO	**1.00**	**0.94**	**0.90**	1.00						
CaO	−0.32	−0.13	0.05	−0.36	1.00					
Na_2O	**0.99**	**0.92**	**0.94**	**0.98**	−0.28	1.00				
K_2O	**1.00**	**0.94**	**0.94**	**0.99**	−0.28	**0.99**	1.00			
TiO_2	**1.00**	**0.94**	**0.91**	**1.00**	−0.34	**0.99**	**0.99**	1.00		
P_2O_5	**0.67**	**0.74**	**0.69**	**0.65**	0.00	**0.62**	**0.64**	**0.68**	1.00	
$SiO_{2\ biog}$	**−0.72**	**−0.69**	**−0.67**	**−0.71**	0.27	**−0.74**	**−0.72**	**−0.74**	**−0.57**	1.00
TOC	−0.50	−0.29	−0.19	−0.53	**0.85**	−0.47	−0.47	−0.51	−0.17	0.40

ICP-AES: inductively coupled plasma–atomic emission spectroscopy; TOC: total organic carbon.

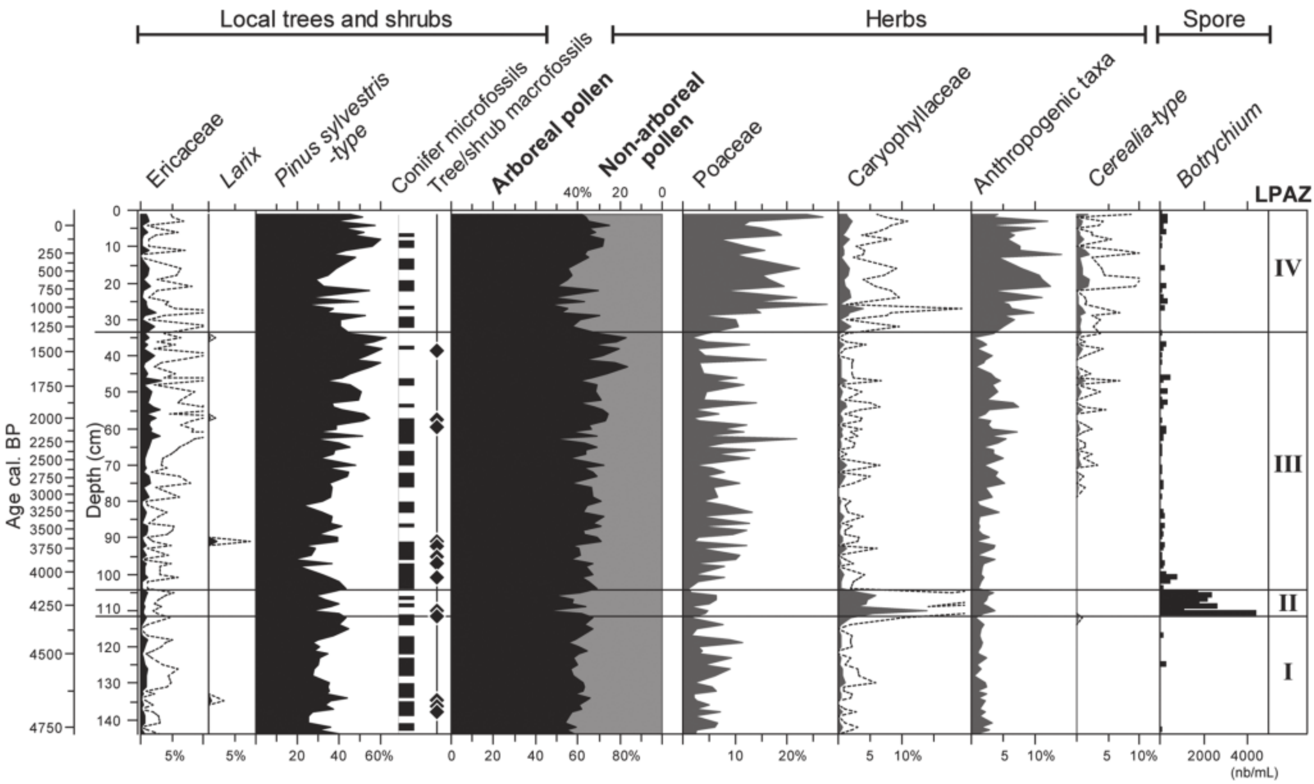

Figure 4. Simplified pollen diagram from core PET09P2. Only local main pollen taxa are represented in relative frequencies. The exaggeration dot line multiplies the relative frequency by a factor of 5. Conifer microfossils represent the occurrence of *Pinus* tracheid and/or conifer stomata. Tree/shrub macrofossil class comprises conifer-tree species as well as indeterminable conifer twigs, barks, needles and leaves (including possible shrubs such as *Juniperus*), and shrubs of Ericaceae.
LPAZ: Local Pollen Assemblage Zone.

catchment as attested by repeated occurrences of conifer tracheid and stomata. Ericaceae, Poaceae and Caryophyllaceae percentages (Figure 4) are low, but continuous and *Botrychium lunaria* spores are present in very low concentrations (*c*. 4400 spores/mL). *B. lunaria* is a fern typical of mountain grassland communities growing on thin, poorly developed siliceous soils. Vegetation around the lake at this time was probably dominated by shrubs and grasses organised in patches composed of Caryophyllaceae (predominance of *Cerastium*), Poaceae, *B. lunaria* and Ericaceae associated with anthropogenic taxa that might indicate the presence of pastoralism or slight clearing of the landscape.

The LPAZ II is characterised by a transitional decrease to 50% of the arboreal pollen, but the frequency of the *P. sylvestris*-type remains high (30%). Ericaceae, Poaceae and anthropogenic taxa show low frequencies. The local vegetation composition remains relatively unchanged relative to the previous LPAZ but Caryophyllaceae and *Botrychium* taxa exhibit exceptionally high values of up to 14% and 4400 spores/mL, respectively (Figure 4).

In the LPAZ III, arboreal pollen remains are dominant, and conifer macrofossils are still present. From 4100 to 2700 cal. BP, an increase of Poaceae to 10% suggests a significant opening up of the vegetation in the upper valley. Whereas the pollen curve of *P. sylvestris* increases, Ericaceae reaches an optimum, contemporaneous with the first occurrences of *Cerealia*, and a rise in the anthropogenic pollen curve and type is recorded.

The LPAZ IV is characterised by a decline in tree taxa to 50% and an increase of Poaceae and Caryophyllaceae. The severe decline of *Pinus* together with the absence of ligneous macrofossils suggests that the catchment has probably been treeless since 1300 cal. BP. Moreover, high frequencies of ruderal–anthropogenic

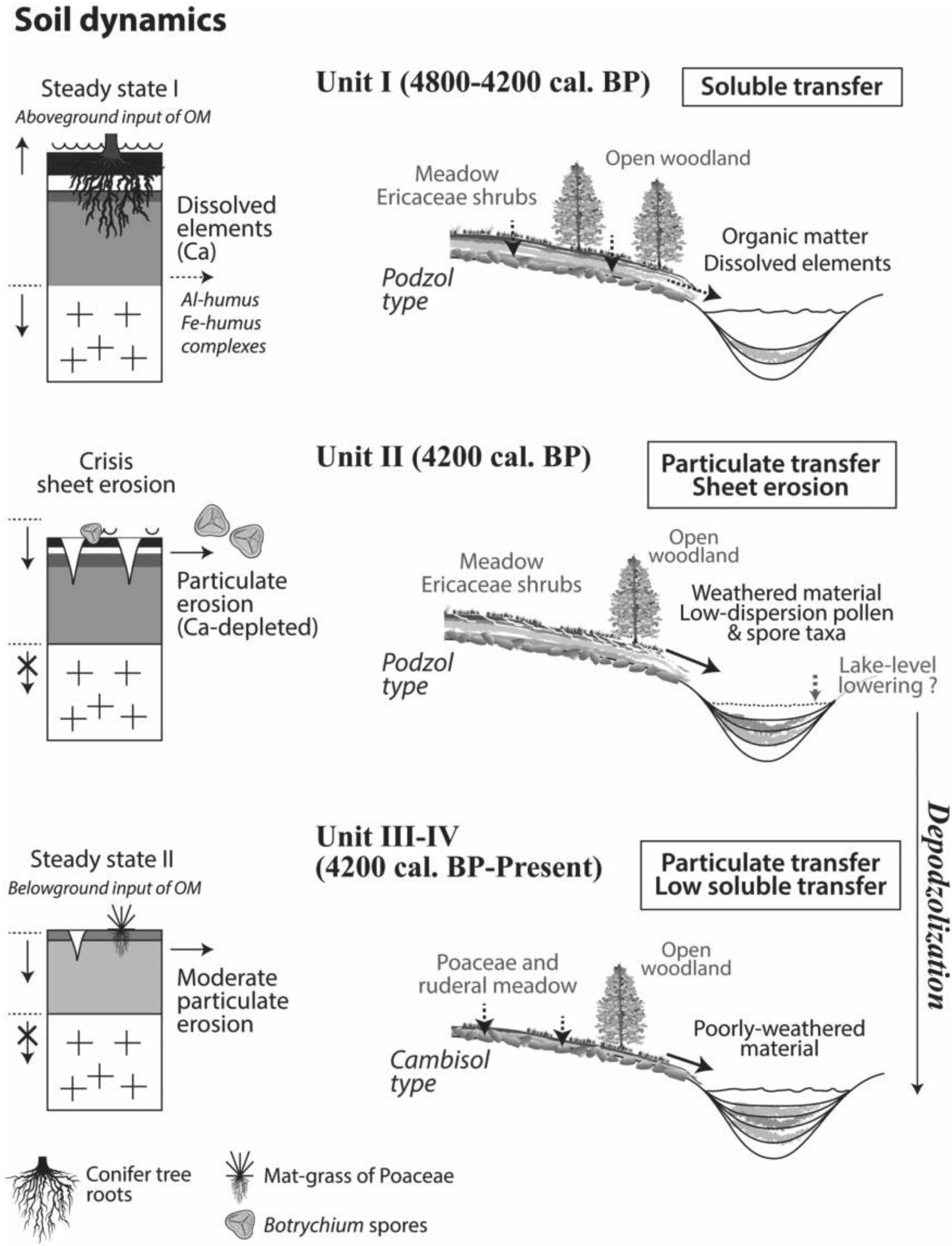

Figure 5. Sketch of evolutionary scenario of soil/vegetation cover interactions recorded in Lake Petit sediments throughout the mid- to late-Holocene period.

grasses confirm the dominance of a grassland environment around the lake, with some markers of nitrogen enrichment of soils, such as *Urtica* (higher than 10%), suggesting intense local grazing.

Discussion

From 4800 to 4200 cal. BP: a mature geosystem on borrowed time

Botanical, sedimentological and geochemical analyses allow us to investigate the relationship between erosional patterns in the Lake Petit catchment and vegetation dynamics over the last 4800 years (Figures 5 and 6). From 4770 to 4350 cal. BP, the local presence of woodlands is attested by a pollen percentage of trees of 60%, by conifer microfossils and by some macrofossils found in the sediment (*Betula* sp. at 142.5 cm in PET09P2). A macrofossil of *Larix* retrieved in coring of riverine peat in another nearby lake, Lake Long (2350 m; Figure 1) has been ^{14}C-dated 3940 ± 30 BP (4510–4260 cal. BP; Suméra and Geist, 2010). The development of larches and pines in association with shrubby patches of Ericaceae (macrofossils in PET09P2 at 135 cm) might have led to the accumulation of a humus layer favouring progressive soil development (Figure 5). In Unit I of Lake Petit, the terrigenous flux is close to zero, suggesting that mobile element enrichment prevailed, with elemental soil loss over this period occurring in

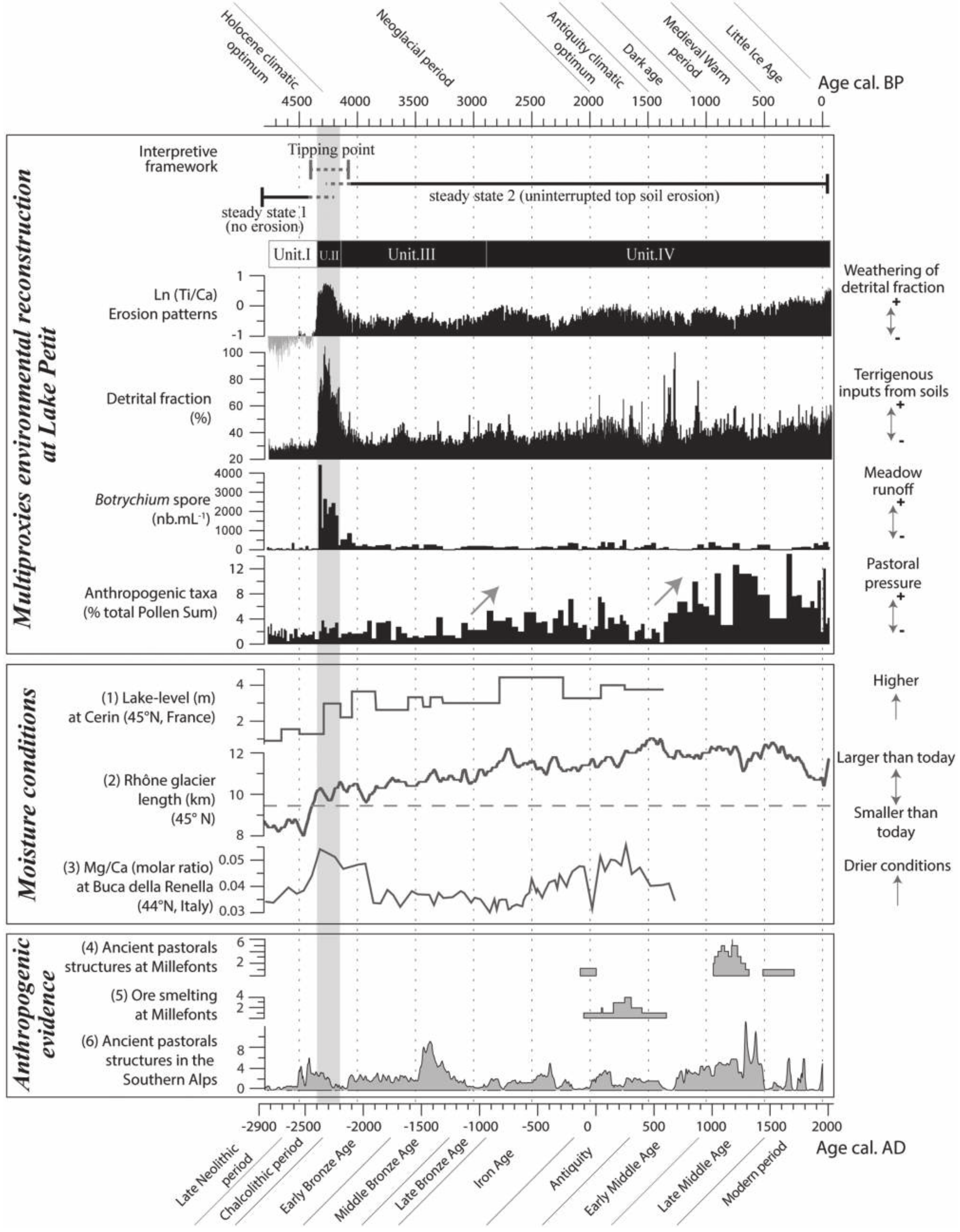

Figure 6. Main geochemical and botanical proxies of core PET09P2 compared with hydrological conditions of (1) Magny et al. (2012a), (2) Goehring et al. (2012) and (3) Drysdale et al. (2006). Number of ^{14}C dates (2σ) in archaeological stratigraphy in the Southern Alps: (4) Suméra et al. (2008), (5) Morin and Rosenthal (2002) and (6) Walsh et al. (2007).

soluble form due to the chemical weathering of primary minerals in the subalpine soil. High diatom abundance (70%) and remarkably high values of TOC (15%) suggest that Si and organic matter were subsequently trapped in the lacustrine ecosystem. Conifer forests on podzols lead to organic matter accumulation and development of mor-type humus, producing organic acid which promotes chemical weathering (Egli et al., 2009). A similar pedogenetic fingerprint in lacustrine sediment has been observed in Lake Thyl and Lake Loup, in the central western Alps (Mourier et al., 2010), where element enrichments (rare earth elements and secondary Al-/Fe-bearing phase in this case) are due to soil acidification during podzolisation of preexisting soils under arolla pine forests, while gyttja accumulated in the lakes.

Multiproxy lake sediment studies covering the Holocene period in high-altitude sites of the Alps highlight a common optimum of soil development with a dominantly conifer-tree forest. Poulenard (2011) has termed this particular phase the 'Holocene Pedogenetic Optimum'. This phase ends at 6500 cal. BP in Lake Thyl (Mourier et al., 2010), 5550 cal. BP in Lake Anterne (Giguet-Covex et al., 2011), 5000 cal. BP in Meidsee (Thevenon et al., 2012), 4500 cal. BP in Lake Loup (Mourier, 2008), 3700 cal. BP in Sägistalsee (Koinig et al., 2003) and 3500 cal. BP in Unterer Landschitzsee (Schmidt et al., 2002; Wick et al., 2003). Mourier (2008) suggests that the maturing and persistence of soils might have been constrained by long-lasting favourable conditions, that is, a warmer climate, low runoff and low anthropogenic pressure. Local bioclimatic and anthropogenic settings and thresholds could explain asynchronous timings of this phase.

From 4200 cal. BP to present: beyond the resilience of the geosystem

Activation of soil erosion and degradation of vegetation cover. The switch from Unit I to Unit II at 4200 cal. BP is recorded by an increasing DF (Figures 5 and 6). Higher values of CIA in Unit II indicate moderate weathering of the parental material. This major change in geochemical regime suggests a drastic modification in pedogenetic processes from progressive development of soils (podzol-type) before 4200 cal. BP to erosion or soil denudation processes. At 4200 cal. BP, runoff, possibly due to more intense precipitation, may have been so frequent as to result in only moderate weathering, thus favouring the development of Cambisols, such as those currently found around Lake Petit. This hypothesis is supported by the simultaneous over-representation of very low-dispersal alpine meadow pollen of Caryophyllaceae and *Botrychium* spores (Figure 6). This type of erosive pattern is rarely identified in pollen records. A high abundance of these taxa concomitant with a detrital pulse attests to erosion of both litter and subsurface soil horizons, probably through sheet erosion processes (Figure 5). However, Lake Grenouilles, close to Lake Petit, also recorded a fall in arboreal pollen together with high percentages of low-dispersal alpine meadow taxa such as Poaceae, Chenopodiaceae and Caryophyllaceae (Kharbouch, 2000). Over this period, a lowering of the percentages of *Larix* may be interpreted as local deforestation favouring the renewed growth of alpine meadows, possibly over-represented as a result of sustained intense runoff on catchment slopes.

The high temporal resolution data obtained at Lake Petit provide evidence of an alpine geosystem capable of tipping rapidly and irretrievably. The system turned from a steady-state without soil erosion (Unit I) to a new state dominated by a degradation trend for both slopes and vegetation cover (Figure 5). From the tipping point (4350 cal. BP), represented by increasing terrigenous inputs, a new sedimentary regime under continuous erosive control has been maintained until today, without any transitional phase of soil stabilisation (uninterrupted soil erosion: Units II, III and IV; Figure 6).

Slopes under anthropogenic pressure? The aforementioned soil erosion trend may be linked to an increase in anthropogenic pressure. Indeed, since *c*. 5000 cal. BP in both the Northern (Curdy, 2007; Krause, 2007) and Southern Alps (Mocci et al., 2008; Walsh et al., 2007), abundant archaeological evidence of repeated seasonal pastoral activities (enclosures for herds with adjacent smaller domestic living areas) has been found at altitudes exceeding 2000 m (Figure 6). Lake Petit recorded between 4770 and 4300 cal. BP, a continuous presence of alpine grasslands characterised by high percentages of ruderal taxa (Figure 6), such as *Rumex* and *Plantago* sp., signifying that local grazing activities occurred in the catchment. As attested by the thousands of carved rocks in the Mount Bego region (20 km distant from Lake Petit), transhumant pastoralism may have started during the Chalcolithic and the ancient Bronze Age (De Lumley and Echassoux, 2009). Although these changes involved human activities and the vegetation cover, no significant soil erosion occurred until 4200 cal. BP. However, starting from 3000 cal. BP (late Bronze Age), intensified pastoral activities appear to have strongly modified the landscape as anthropogenic grasslands, together with Ericaceae shrublands, progressively developed in association with poorly developed soils (Figure 5). At this time, the detrital input increased slightly. Finally, a rapid development of anthropogenic grassland communities between 1800 cal. BP and 500 cal. BP attests to a local upswing in human pressures on the Lake Petit catchment. Land-use changes were not accompanied by a significant increase in TOC concentrations (nor in TOC flux), which could have been an effect of long-lasting depletion of soil organic carbon. Further analyses on nutrient availability will be needed to verify this hypothesis. Maximum human pressures indicated by pollen coincide with evidence of pastoral and mining activities in the Lake Petit catchment. Ongoing ^{14}C investigations on archaeological remains indicate mining activity between 2200 and 1600 cal. BP (Morin and Rosenthal, 2002; Pagès, 2009), and excavated stratigraphies on pastoral enclosures attest to the presence of shepherds and livestock between 1000 and 400 cal. BP (Suméra et al., 2008). As a result, from 3000 cal. BP to present, soils lost protection from the arboreal cover, and were probably (1) compacted by domestic livestock and (2) recurrently eroded by runoff intensified by general deforestation for pastoral and mining purposes. Regarding this period, it must be noted that soil erosion remained regular but limited compared to the detrital pulse at 4200 cal. BP, although human disturbances had intensified. Therefore, even though human activities have had a direct impact on persistent sediment availability at least since 3000 cal. BP, human pressures do not explain the sharp change in soil cover at 4200 cal. BP.

Increasing erosivity and sediment transfer under climatic control?

At the millennium scale: influence of Neoglacial climatic change. At *c*. 4000 cal. BP (corresponding to the onset of the so-called Neoglacial period), there is a general agreement in the climatic archives regarding an increase in humidity and cooling temperatures as shown by glacier advance (Goehring et al., 2012; Holzhauser et al., 2005) and treeline regression (Nicolussi et al., 2005) throughout the Alps and the northern Mediterranean (Zanchetta et al., 2012). Water-levels in many lowland lakes of western Europe (Lake Cerin, Magny et al., 2012a; Lake Bourget, Magny et al., 2012b) and in the northern Mediterranean (Lake Saint Léger, Digerfeldt et al., 1997; Lake Accesa and Lake Ledro, Magny et al., 2012a) exhibit a slight transition from low stands before ~4200 cal. BP to higher levels (Figure 6) thereafter, which indeed could suggest a common climatic forcing factor towards wetter conditions at *c*. 4200 cal. BP. Orbital changes involving both a general gradual decrease in summer insolation and a major reversal in insolation seasonality (Berger and Loutre, 1991) might have led to a more southward position of the Westerlies (Magny et al., 2011) and to wetter conditions over Europe (Andresen and Björck, 2005; Hughes et al., 2000; Larsen et al., 2012). These climatic changes have been used to subdivide the Holocene into mid- and late Holocene (Walker et al., 2012).

The expression of this long-term regional-scale climatic influence might be more contrasted in alpine Mediterranean highlands where climatology is characterised by high inter- and intra-annual variability in precipitation. Situated at 1308 m in the Southern Alps, Lake Saint Léger has recorded a lowering of its level since 4100 cal. BP, reaching a lowest level between about 3500 cal. BP and 2600 cal. BP (Digerfeldt et al., 1997). According to the synthesis of Muller et al. (2012), many lakes in the Southern Alps (ranging from 950 to 1800 m a.s.l.) experienced complete drying out between 6600 and 2500 cal. BP. This trend has been due to high autochthonous organic production and possibly seasonal changes towards drier conditions in winter and/or warmer conditions in summer. Arguments in favour of seasonal change are also supported by a strong decrease in sedimentation rates in fluvial systems of the Southern Alps (Miramont et al., 2008).

At Lake Petit, the bipartition from the mid- to late Holocene can be clearly identified by soil cover modification and linked with a change in precipitation regime. Therefore, it can be assumed that wetter conditions might have played a major role in frequent sediment transfers to the lake. However, can a progressive change in the rainfall regime explain the state of no return of extremely sensitive soils?

The 4.2 ka event: an environmental crisis at Lake Petit? An abrupt climate pulse and/or rapid environmental changes have been identified around 4200 cal. BP in the Northern Hemisphere (Booth et al., 2005; Huang et al., 2011; Magny et al., 2009, 2012b; Staubwasser et al., 2003) and regionally in the northern part of the Mediterranean basin (Bruneton et al., 2002; Drysdale et al., 2006; Miramont et al., 2008). This event could correspond to the over-represented alpine meadow pollen zone of Lake Grenouilles according to an earlier ^{14}C age of 4700 ± 60 BP (5310–5580 cal. BP) obtained by Kharbouch (2000). At Millefonts, the transition from stable soils (Unit I) to permanent soil erosion (Units II, III and IV) is expressed by a sharp increase in terrigenous inputs at 4300 cal. BP. According to the age–depth model (Figure 2), this detrital pulse occurred between 4430 and 4055 cal. BP (2σ interval), which means that the event duration did not exceed 375 years. This detrital event profoundly increased slope sensitivity to erosion processes compared to the impact of the period of increasing human pressure (from 3000 cal. BP to present and in particular during the Mediaeval Period). Consequently, the hypothesis of a local but intense human impact during a short period is unconvincing to explain alone the severity of soil destabilisation. Besides, the speleothem record of the Buca della Renella cave in mid-latitude Italy (Drysdale et al., 2006 and Figure 6) indicates that a severe drought occurred between *c*. 4100 and 3800 cal. BP. The pollen diagram of Lake Petit does not show any significant evidence for a strong reversal in vegetation dynamics that could indicate a direct response to reduced moisture. Increase in dry density and increasing soil erosion could be indirect responses to long-lasting drought conditions in the catchment, which may have led to lake-level lowering during periods of earlier spring snowmelting. Hydrological changes could have been a direct response to this abrupt climate event. This hypothesis requires testing in the future. Finally, smooth changes in precipitation regime do not explain more satisfactorily the 'environmental crisis' signature recorded at 4200 cal. BP. The abrupt pulse recorded in Lake Petit could indirectly correspond to a 'rapid' climatic oscillation, within a background of both 'smooth' anthropogenic pressure and 'smooth' orbitally induced climate change at the mid- to late-Holocene transition.

Conclusion

For the first time in the southern French Alps, a high-resolution multiproxy investigation involving sedimentological, geochemical and botanical data has enabled a detailed 5000-year reconstruction of the dynamics of a fragile alpine ecosystem. At Lake Petit, the outstanding quality of the diatomite-rich sedimentary archive was a major advantage for tracking any detrital pulse and/or regime that occurred during the second half of the Holocene. Therefore, four main periods have been identified.

The period from 4800 to 4200 cal. BP is characterised by low physical weathering of silicates and presence of conifer woodlands, while pure diatomite was deposited in the lake. This stable phase of progressive soil development is characteristic of the Holocene Pedogenetic Optimum. An abrupt detrital event at 4200 cal. BP could be the indirect response of the local environment to the rapid climate oscillation of the 4.2 kyr event. The period from 4200 to 3000 cal. BP following this abrupt detrital event was marked by long-lasting soil erosion and anthropogenic disturbances, leading to a major pedogenetic process reversal towards poorly weathered soils. The period from 3000 cal. BP to present was characterised by peak anthropogenic pressure. The erosion dynamic induced at the onset of this phase has been continuous, without any transitory phase of soil stabilisation, and it still prevails today.

We have shown that the degradation of such an environment might have been more complex than hitherto assumed. The dynamics of the Lake Petit catchment were not driven by long-lasting disturbances but possibly by concomitant rapid changes in climate and smooth anthropogenic disturbance. Indeed, as the threshold of resilience was reached at 4200 cal. BP, the soil around the lake seemed to have become progressively transformed while human activities were on the rise. Over the last 2600 years (around 4000–1300 cal. BP), this fragile equilibrium between human-induced erosion and the persistence of degraded soils might have lasted before being terminated undoubtedly during the Middle Ages. From that period, anthropogenic pressure reached its maximum in the vicinity of the lake, inducing the last recorded stage of degradation of the environment.

Therefore, since 4200 cal. BP, the breakdown occurred in two steps: the first rapid phase (possibly climate driven) triggered weaknesses that made the system more vulnerable to agro-pastoral pressure which occurred during the second phase. Moreover, since this tipping point, the geosystem might have no longer been subjected to a seasonally contrasted Mediterranean climate regime. As a result, Lake Petit history illustrates clearly how the response of a geosystem facing both climate and human impact could be non-linear. The debatable question of the 'weight' of the human-triggered soil erosion after 4200 cal. BP can be broadened to include the current problem of global change and the role of human activities, at both local and global scales. Is climate change threatening to the resilience of environments? Is human impact weakening these environments in the face of climate change? Both questions might find answers in sediment archives such as those of Lake Petit.

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