Lake Ohrid, Albania, provides an exceptional multi-proxy record of environmental changes during the last glacial–interglacial cycle

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A B S T R A C T

Multi-proxy analyses on core JO2004-1 recovered from Lake Ohrid (40°55.000 N, 20°40.2976, 705 m a.s.l.) provide the first environmental and climate reconstruction in a mountainous area in Southern Europe over the last 140,000 years. The response of both lacustrine and terrestrial environments to climate change has been amplified by the peculiar geomorphological and hydrological setting, with a steep altitudinal gradient in the catchment and a karstic system feeding the lake. The karstic system was active during interglacials, leading to high carbonate production in the lake, and blocked during glacials as a result of extremely cold climate conditions with permafrost in the mountains. At the Riss-Eemian transition (Termination 2) the increase in lacustrine productivity predated forest expansion by about 10,000 years. In contrast, the Late Glacial–Holocene transition (Termination 1) was characterized by the dramatic impact of the Younger Dryas, which initially prevented interglacial carbonate production and delayed its maximum until the mid-Holocene. In contrast, forest expansion was progressive, starting as early as ca. 38,000 ago. The proximity of high mountains and the probable moderating lake effect on local climate conditions promoted forest expansion, and contributed to the make the surroundings of Lake Ohrid favourable to forest refugia during the last glacial, usually steppic, period. Our study of sedimentology, mineralogy, geochemistry, magnetics, paleontology and isotopes illustrates the non-linear response of terrestrial and lacustrine ecosystems to similar climate events, and demonstrates the potential of Lake Ohrid as an excellent paleoarchaeological archive during the Quaternary.

1. Introduction

Lake Ohrid, Albania, is an exceptional site for reconstruction of regional climate history over repeated glacial-interglacial fluctuations and its impact on terrestrial and lacustrine ecosystems. Not only the largest southern-European freshwater body (355 km², 289 m depth), it is likely the richest in Europe and renowned for high level of endemism, with at least 203 fish species described (Stanićov, 1960). The exceptional thickness of sedimentary deposits (Dumardanou et al., 2005) makes it directly comparable to the Mediterranean lowland site of Tanaghi Philippin in NE Greece which provides a continuous 1.35-million-year pollen record (Reedkis et al., 2006). Its location at middle altitudes in a rift basin surrounded by high escarpments within a key area at the confluence of central-European and Mediterranean climatic influences, provides a unique opportunity to study the impact of climate changes on middle- to high-altitude forest ecosystems. Particularly, Lake Ohrid is an exception location which allows testing of the "glacial refugia" hypothesis for southern Europe bordering the Mediterranean (e.g., Drėžė et al., 2000; Tzedakis et al., 2002). Previous studies of the Lake Ohrid sediment record focused on palaeoecological and sedimentological aspects during a 40,000 years period (Reedkis and Kilham, 1983; Wagner et al., 2008a,b). Here we present for the first time a record of
environmental and climate change over the last 140,000 years. Attention is paid to selected proxies to discuss the influence of climate change on the karst system and lake hydrology, as well as on regional vegetation response to glacial-interglacial variations.

2. Modern setting

Lake Ohrid lies in a strongly asymmetric, N-S oriented half-graben at the Macedonia-Albania border. It is bounded by faults running N to NNE which affect, to the north and east, carbonate rocks of Triassic and Jurassic age, and ophiolitic rocks of Jurassic age to the south-west (Fig. 1). The southern end of the basin connects with a small graben filled by Pliocene continental mudstones and sandstones, overlain by fluviolacustrine sediments of Holocene age (Nicol and Chardon, 1981). The basin is filled by several hundred meters of sediments deposited since around 85,000 yr BP (Dumitracov et al., 2005). The modern lake is holomictic. Today roughly half of its water is derived from a number of springs located in the SE part of the lake, draining a karstic system which, in turn, is fed by water from nearby lake Prespa and infiltration of rainwater (700 mm/yr on average) in the Galicica mountain range. The remaining water comes from rivers (e.g., the Struga River to the north) and direct precipitation. A single outlet (the Black Drin River) to the

Fig. 1. Geological/geomorphological map of the Lake Ohrid basin, southeastern Balkans, showing the location of the core site.
north and significant evaporation (estimated at 145 mm/yr) complete the water budget (Matzinger et al., 2006). The small size of the Lake Ohrid drainage basin (2600 km²) and its large volume (55.4 km³) lead to a water residence time of about 70 yrs (Matzinger et al., 2006). The catchment area covers a large altitudinal range, from 790 m to more than 2200 m. The mean annual temperature is 11.5°C, with winter temperatures varying from −2.3°C to 6.6°C and summer temperatures from 10.5°C to 22.3°C. Precipitation is principally during winter and spring. As a result of the high topography, the vegetation is distributed in altitudinal belts with mixed deciduous forests including Carpinus orientalis, Quercus troiana, O. frainetto, O. cerris at lake level followed by Fagus moesiaca, Alnus alba, A. horrida-regni at the upper limit of the forest, and sub-alpine grassland with Juniperus macrocarpa above 1800 m in the Mavro Tourlakos and Mountainous areas.

3. Materials and methods

3.1. Core J02004-1

Core J02004-1 consists of two series of core sections (each up to 3 m long) recovered ca. 5 m apart from a single site in the southwestern part of the lake (40°55.930 N, 20°40.297 E, 705 m a.s.l.). The uppermost ca. 10 m of sediment were recovered in consecutive 3-m long sections using a hammer-driven Niedermeier piston corer of 63 mm diameter. To obtain a continuous sedimentary record four sections were recovered from the first hole (J02004-1), and three sections with a planned depth offset of 1.5 m from the second hole (J02004-1a). The overlapping sections were visually cross-correlated using marker layers clearly identified in both sequences, and the resulting composite profile was checked for consistency using the magnetic susceptibility record. The depth information used in the paper refers to a composite linear master depth scale (MCD). In addition to the core sediments, 15 samples taken from a variety of rocks and sediments outcropping in the catchment basin of Lake Ohrid have been studied for mineralogical investigations of the bulk and clay fractions and geochemical analyses.

3.2. Magnetic susceptibility

Core sections were sub-sampled with U-channels at the core repository of the Laboratoire des Sciences du Climat et de l’Environnement (LSCE) in Gif-sur-Yvette (France). The volume low-field susceptibility (χv) was measured at a high resolution. χv is determined by the amount of ferroferri magnetic grains, but mineralogical changes and grain size variations may also influence the susceptibility signal. Paramagnetic material (clay) and diamagnetism also contribute to χv when the magnetic grain content is low. Measurements of the low-field bulk susceptibility (χv) were performed using a small-diameter Bartington sensor loop mounted in line with a track system designed for a-channel.

3.3. Sedimentology, mineralogy and sediment geochemistry

Sediment fabrics were visually described on bulk samples. In addition, one hundred and ninety samples were analysed for bulk and clay mineral investigations. Each sample was divided into two sub-samples: one sub-sample was crushed in a grinder and pressed into a holder for bulk mineral analyses. The other sub-sample was washed on a 63 μm mesh sieve. The <63 μm fraction was decalcified using a solution of 10% HCl, rinsed with de-ionized water, and decarbonized through repeated centrifugation. The clay fraction (<2 μm) was decanted and deposited onto a glass slide. X-ray diffraction (XRD) analyses were conducted on natural clay slides after ethylene-glycol solvation and heating at 430°C for 2 h. Determination of percent bulk and clay minerals was based on the respective XRD peak areas using MacDiff version 3 software (Petschick, 1998; relative error is ±5%.

Samples subjected to bulk geochemical analyses were processed at UMR 5538 “Domaines Océaniques”, IUEM-UBO, Brest, France. Major elements were measured by ICP-AES with an ISA Jobin-Yvon JY 70 Plus apparatus. Calibration were checked using the GIT-IFGW (Groupe International de Travail-International Working Group) BE-N, WS-E, PM-S, AC-E and the CRMP (Canadian Certified Reference Materials Project) LPSD-1 international standards. Trace element and rare earth element (REE) measurements were conducted by ICP-MS using a Finnigan Element 2 ICP-MS. Results obtained for the CRMP (Canadian Certified Reference Materials Project) LPSD-1 were reproducible with precision for trace elements <5%, except for Zn and Hf which were not used to characterize the samples.

3.4. Dating procedures

3.4.1. Radiocarbon measurements

Seven AMS radiocarbon measurements were performed on terrestrial organic matter and charcoal remains. The 14C activity was determined by UMS-ARTEMIS (Pelletron 3MV) AMS Facilities. Raw 14C dates were converted to calibrated ages using Calib 4.4.3 (Stuiver et al., 2005) except for the oldest radiocarbon age, for which polynomial equations (Bard, 1998) have been used.

3.4.2. Tephrochronology

For tephrochronology, the core was continuously screened at 1-cm intervals for volcanic particles (glass shards, mafic crystals, volcanic rocks). The sediments were washed and sieved at 125 μm and 40 μm using distilled water. The sediment fraction less than 40 μm was collected using a paper filter and progressively dried through a water pump. The sediment fraction 40–125 μm was carefully inspected under stereo microscope. Chemical analyses on glass shards from tephas were performed at the Dipartimento di Scienze della Terra (University of Pisa), using an EDAX-IXX micro-analyser (EDS analyses) mounted on a Phillips SEM 515 (operating conditions: 20 kV acceleration voltage, 100 s live time counting. 10−8 Am beam current. ZAF correction). Instrument calibration and performance are described in Mariandini and Straneo (1998). The major-element chemistry of crystal-free glass shards were used to classify the individual tephas layers and correlate them to different archives. SEM back-scattering images were used to describe the texture of the ground mass and to select micro-crystals for micro-analyses.

3.5. Pollen analysis

One hundred and fifty five samples were taken at 1 to 15-cm intervals for pollen analyses. Samples were processed using the standard HF method (Faegri and Iversen, 1989) and sieved on a 5-μm mesh. Addition of a known amount of exotic makers (Lycopodium) allowed calculation of pollen concentration (grains per gram of dry sediment) and influx values (grains per cm² per year). In total, 110 different pollen types were identified for a mean pollen sum of 450 grains per sample (maximum count = 1480 grains). Pollen data are expressed as percentages calculated against the sum of all the determined pollen types except aquatics, spores (bryophytes and ferns) and damaged grains. Pollen grains were determined using regional pollen atlases from Europe and the Mediterranean (Reille, 1952; Chester and Raine, 2001).

3.6. Ostracod isotope studies

Ostracod shells were extracted from 1-cm sediment slices taken at 5-cm intervals throughout the core, with the exception of a few continuously sampled sections at major lithostratigraphic transitions (Belchev et al., 2006). For sample preparation, the sediment was disaggregated with H2O2, and gently wet-sieved. The >125 μm fraction was washed with ethanol and dried at room temperature. Ostracod shells were extracted and cleaned. Ostracod species identification
follows Klie (1936a,b; 1942) and Miltiadj (1961). The stable-isotope composition of the shells was analyzed at the Leibniz Laboratory in Kiel using a Finnigan MAT 251 mass spectrometer equipped with an automatic carbo-Kiel CO₂ preparation device. Isotope ratios are given relative to the PDB standard in the usual notation with a total accuracy on the order of 0.35 % and 0.05 % for δ¹³C and δ¹⁸O, respectively.

4. Data (Fig. 2)

4.1. The sedimentary record

4.1.1. The present-day catchment basin of Lake Ohrid

Mineralogical investigations of the bulk and clay fractions have been conducted on a variety of rocks and sediments occurring in the catchment basin of Lake Ohrid (Table 1). Lizardite and dioctahedral mica are the major constituents of the ophiolites, which weather primarily into smectite. The bulk composition of mudstone deposits, Neogene and Quaternary sediments are largely dominated by quartz, associated with feldspars, clay minerals and, locally, a proportion of ophiolite minerals (lizardite and chlorite). Clay minerals include illite-smectite and chlorite-smectite types, illite, smectite and kaolinite. However, the relative proportions of the minerals vary strongly depending on the nature of the substrates in the different parts of the drainage basin. In addition, irregular mixed-layered clays and illite-smectite type contain a high proportion of illite layers and probably reflect subsequent weathering of illite in the porous terrigenous sediments. The carbonates that outcrop in the drainage basin consist of calcite but do not contain other types of minerals and are not weathered into clay minerals. It is remarkable that calcite is absent or occurs in very minor abundances of less than 5% in molasse deposits and other terrigenous sediments of Pliocene and Quaternary age (Fig. 2).

Geochemically, the ophiolitic formations are dominated by SiO₂ and MgO, with minor amounts of Fe₂O₃tot. Al₂O₃. The Triassic-Jurassic carbonates are purely formed by Mg-calcrete (56 wt.% CaO and 0.45 wt.% MgO and the continental mudstones and sandstones of Pliocene age at the southern end of the Ohrid basin are characterized by Al₂O₃ and CaO, with minor proportions of Fe₂O₃tot and K₂O.

4.1.2. Core 9502/D04-1

The sediment series generally includes calcareous and siliciclastic elements. Calcareous elements occur primarily in the form of clay-sized to sand-sized aggregates and biogenic remains (ostracods), which are especially abundant from the top of the core to 75 cm (calcareous clay and sand in the calcarcous mud) and from 470 to 924 cm (calcareous clay to sand in the calcarcous mud). Siliciclastic elements are especially abundant from 75 to 470 cm (clay, silty clay and sand) and from 524 cm to the bottom of the core (silty clay). Sandy intervals at 188 cm and 240–246 cm are conspicuous tephra layers. The sediment color ranges from beige to grey, the darkest colors being associated with occurrences of organic remains, principally plant debris.

Mineralogical assemblages of the core’s bulk sediment are dominated by quartz and calcite. Quartz is associated with smaller proportions of feldspars and clay minerals. Similar, quartz dominated

<table>
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<th>Table 1</th>
<th>Mineralogical and geochemical composition of the main geological outcrops in the Ohrid-Maliq basin.</th>
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mineragical assemblages are also found in alluvial deposits and other Pleistocene and Quaternary sediments of the region. Mineral particles are typically terrigenous and originate from erosion from the adjacent drainage basin by running waters. However, feldspars significantly increase at several intervals, where they may result from in-situ alteration of tephra layers. Other terrigenous elements include pyroxenes and ophiolite minerals (principally lizardite and clinochlore), derived from ophiolite formations which largely outcrop in the drainage basin. When present calcite may represent up to 90% of the bulk sediment. Calcite occurs as biogenic elements (ostracods principally) or silt-sized to clay-sized aggregates.

Clay mineral assemblages are dominated by illite, which occurs in significant amounts in alluvial deposits and other terrigenous sediments of the drainage basin. In both cases, illite is associated to variable percentages of chlorite, kaolinite, and random mixed-layered clays. The association of minerals derived from poorly weathered substrates (illite, chlorite, random mixed-layered clays) and from areas of intense chemical weathering (kaolinite) probably results from intense erosion of soils and substrates in adjacent mountain areas by run-off. It is noteworthy that maximum abundances of illite and associated minerals coincide with occurrences of calcite. Maximum abundances in illite and chlorite alternate with intervals of abundant smectite. In the modern world, abundant smectite is often found in warm areas of seasonal precipitation (Chamley, 1989; Weaver, 1989). Its formation is strongly enhanced on talus substrates, and small proportions of the mineral have even been described in the weathering profiles of basaltic substrates of cold areas, for example in modern West Antarctica (Campbell and Claridge, 1998). Ophiolite outcrops of the drainage basin of Lake Ohrid have shown how weather into smectite. Maximum abundances of smectite coincide with maximum contents of clay minerals and quartz in the bulk sediment, indicating significant development of weathering and erosion in the catchment.

4.2. Datasets
4.2.1. Radiocarbon measurements
Radiocarbon measurements (Table 2) from the uppermost part of the core are in chronological order from 20.5 cm to 40.4 cm, whereas it was impossible to obtain sufficient organic matter for reliable dating below 100.5 cm. The age of 30, 100 ± 1,200 14C yr BP (207.6 cm) is therefore only indicative.

4.2.2. Tephra
Volcanic material was recognised at 188, 240-246, 575-576 cm and 939-941 cm (Table 3). Two of these tephra layers (at 188 cm and
240–246 cm) were visible as discrete sandy-silty layers, whereas the other two are crypto-tephra (575–576 cm and 939–941 cm) mixed with lake and clastic sediments. A detailed description of the tephra layers of the studied core sequence and in-depth discussion of their correlation and origin can be found in Caron et al. (submitted for publication). The chemical composition of glass shards and microscopic fragments permitted the correlation of Lake Ohrid samples to other tephra layers already recognized in the Mediterranean basin: the tephra layer at 188 cm has been correlated with the Y3 tephra layer dated at 306.7 ± 2.0 yr BP (Keller et al., 1978; Zanchetta et al., 2008). The base of the tephra at 340–246 cm has been correlated with the Y5/Campanian Ignimbrite tephra layer, widely recognized in the central-eastern Mediterranean and Eastern Europe mainland (Keller et al., 1978; Pyle et al., 2006; Giaccio et al., 2008; Aksu et al., 2008; Calanchi and Dinelli, 2008) and dated to 638 ± 110 yr BP (De Vries et al., 2001). The crypto-terphra material at 575–576 cm depth is correlated with the X6 tephra layer (Keller et al., 1978) on the basis of the good compositional match with analyses from Lake Grande di Monticchio (Wulf et al., 2006) and the C-31 marine tephra layer (Brauer et al., 2007; Patrone et al., 2008) (Fig. 3). The 234Th/230Th age of 170,000 ± 2000 yr BP for the X6 tephra (Keller et al., 1978) is in good agreement with the suggested age of 108,430 ± 170 yr BP obtained from the varve-supported Monticchio chronology and the suggested age of C-31 tephra layer at 167,000 yr BP (Patrone et al., 2008). The volcanic fragments of the crypto-terphra at 939–941 cm depth are a Panamite (Givet et al., 1984, 1988). In marine cores from the Mediterranean (Patrone et al., 2008), two tephra layers match the lithology and composition of this tephra layer, namely P11 (113,000 yr BP) and P12 (140,000 yr BP). Both tephra layers originate from Pantelleria Island, and correspond to the ignimbrite P1 (dated at 133,100 ± 3300 yr BP) and the Waldorf Tuff's (dated at ca. 135,000 yr BP), respectively (Mahood and Hildreth, 1986). Stratigraphic considerations support correlation of the 941–941 cm tephra layer with the P11 marine tephra layer, and thus with ignimbrite P4 deposition on Pantelleria. The tephra at 943–941 cm is located within the rising calcite content and increasing arboreal pollen values which unequivocally indicate that this tephra was deposited close to the inception of the Last Interglacial. Although the onset of the Last Interglacial may have been recorded at different times in different archives (e.g., Sánchez-Gotli et al., 2002), a putative age of 144,000 yr BP (which correspond to P12 tephra layer) is several ten of thousands of years too old for the inception of the Last Interglacial around the Mediterranean (e.g., Sánchez-Gotli et al., 1999, 2005).

### 4.3 Pollen

The Ohrid pollen record shows the classical opposition between arboreal pollen (AP) and non-arboreal pollen (NAP) percentages throughout the last climatic cycle (Bond et al., 2001). AP pollen dominate with percentages higher than 80% between 879 and 597 cm, 567 and 557 cm at 477 cm and then from 1093 cm to the top of the sequence. A minor peak of 77% is recorded between 335 and 280 cm. Trees and shrubs are mainly Pinus (maximum = 73%), Abies (54%), Quercus (30%), Carpinus (14%), Juniperus (13%), Quercus ilex (12%), Hippophae rhamnoides (8.6%), Tilia (7.7%), Fagus (7%), Betula (4.6%), Olea (3%), and Corylus (1.7%) recording distinct influences from Mediterranean and temperate mid- and high-elevation forests. Three pollen types of steppe origin dominate the herbaceous plant communities (Asteraceae, Poaceae, and Chenopodiaceae) with maximum percentages of 46% (28%), 30% (213 cm) and 15% (516.9 cm), respectively.

In detail, changes in forest vegetation record the spread of Pinus and Quercus robur, associated, to a lesser extent, with Quercus ilex and Tilia during the first phase of forest expansion (1789–791 cm). These taxa strongly decline (Pinus, Q. robur, Q. ilex) or disappear (Tilia) to the benefit of Abies which massively expands (791–506.9 cm), in association with Carpinus (peak reached at 737.5 cm) then with Picea (peak reached at 665.9 cm). Abies and Pinus strongly increase during the following forest phases at levels centred at 556.9 cm and 476.9 cm, then between 1255 and 545.3 cm in association with Q. robur. The forest phase recorded between 212.5 and 270.0 cm displays a slightly different pattern with Pinus dominating the pollen assemblage and Q. robur, Picea and Abies reaching 10% or less, only.

### 4.4 Ostracods

The lowest part of the core, from the base to 640 cm, is devoid of ostracod valves. At 940 cm the ostracod valves appear and reach a maximum of high species diversity at 880 cm. Ostracod valves are well preserved and taxonomically diversified during the hole interval from 940 cm to 473 cm. No ostracod valves are found during the following interval from 473 cm to 120 cm, apart from some broken valves found in a short 16-cm-thick interval between 320 cm and 340 cm. At 120 cm, we record the re-occurrence of the ostracod valves which are present until the top of the core.

The mean conodont species 87Sr/86Sr record starts with values around 0.7095 at 940 cm followed by an increasing trend which continues until 855 cm reaching a maximum of 2%. After 890 cm, 87Sr/86Sr decreases rapidly to 0.7095 and drop to a minimum of ~2% around 780 cm. The
period between 750 cm and 550 cm is marked by two consecutive positive excursions, with a first maximum of 2% at 570 cm and a second maximum of 0.5% at 550 cm. The period after the hiatus is marked by a single positive excursion starting with -1% at 530 cm, increasing to 1.2% between 520 cm and 480 cm to end with -1% just after 480 cm. The crenoid valve fragments found in the short 10 cm interval at 340 cm show high δ^13C values around 1% close to the values before 890 cm and the positive excursion around 750 cm and 550 cm. In the uppermost part of the record, δ^13C values of the crenoid valves start with moderately low values between 0% and -1% at 110 cm. However, during this period ostracod mean δ^13C values scatter around -0.5%, interrupted by a very short excursion to very low δ^13C values below -3% around 70 cm.

5. Environmental reconstruction

5.1. A sedimentological record controlled by a karstic system

The most striking characteristic of the Lake Ohrid sediment sequence is the strong variation in calcite, which dominates from 870 to 590 cm and from 800 to 100 cm, and is absent from other intervals (Fig. 2). When present, calcite may represent up to 90% of the bulk sediment. Besides biogenic remains such as ostracod shells (not exceeding more than 0.1% of the sample mass), calcite principally occurs as angular to sub-angular, generally silt-sized particles. Calcite is either absent, or only occurs as trace amounts of less than 5% in alluvial and mollusca sediments that outcrop in the adjacent drainage basin. In addition, sandy intervals of the core which may indicate some degree of reworking from shallower, nearshore areas of the lake which are closer to limestone outcrops, generally contain smaller proportions of calcite than adjacent all and clay sediments. Therefore, we consider the contribution of terrigenous calcite as limited. Calcite aggregates of similar morphologies have been described in modern low salinity to fresh water environments of temperate and tropical latitudes (Stabel, 1982; Talbot and Allen, 1996) saturated in CO_2^- (Fugler and Kelts, 1983) where they are considered of authigenic origin. Similar calcite aggregates also characterize Holocene paleokarst deposits of Yemen where matching carbon isotope discrimination in bivalve shells and bulk sediment suggest coeval in-situ formation of biogenic and abiotic calcite (Lézine et al., 1998). Based on identical mineralogy, morphology, and occurrence of the latter calcite aggregates, we consider calcite particles in Lake Ohrid sediments to be of predominantly authigenic origin. Limestone outcrops in the drainage basin of Lake Ohrid are probably dissolved by rain water and soil acids. Calcium and carbonate ions are transported to the lake by run-off and via the karst aquifer. There, carbonate/bicarbonate reactions and photosynthesis progressively decrease the availability of free hydrogen ions and raise the pH. Dissolved carbonate and calcium ions are used by aquatic biota (ostracods, molluscs) to build their mineral parts, or precipitate directly in the epilimnion when at high concentrations, for example during spring and summer algae blooms. At long timescales, calcite formation may either occur during intervals of high precipitation when transport of dissolved calcium and carbonate ions to the lake increase, high evaporation, which increases their concentration in the water, and/ or elevated temperature, which lowers the saturation threshold for calcite. These conditions most likely occur during interglacials. Conversely, calcite deposition is hampered under dry/dry cold climatic conditions due to decreased influence of running waters and the cessation of the karstic flow. In addition, cold conditions increase the capacity of lake waters to absorb carbon dioxide and reduce photosynthesis. This in turn fosters the concentration of hydrogen ions and decreases the pH of the lake waters. Such conditions likely correspond to glacial intervals.

In the absence of dilution by calcite particles, siliciclastics dominate during cold/dry intervals. Siliciclastic assemblages are diversified. They include typically terrigenous elements such as quartz, feldspars, clay minerals (mainly illite, associated with variable percentages of chlorite, kaolinite, random mixed-layered clays and smectite), pyroxenes, and opaque minerals (i.e. ilmenite and chlorite). Lacustrine siliciclastic assemblages are very similar in composition to those found in mollusca and alluvial deposits, suggesting that they are derived from the erosion of the catchment basin by running waters. The presence of clay minerals of diverse origins indicates coeval formation of soils, sediments and substrates. It is remarkable that maximum abundances in illite and chlorite coincide with intervals of highest calcite contents, indicating that strong precipitation also favoured an intense erosion of poorly weathered substrates in the drainage basin of the lake. Intervals of increased smectite contents (which coincide with low calcite and high quartz contents) suggest that dry/cold climatic conditions nevertheless allowed some degree of chemical weathering of the opaquesubstrates. Such a transient interval of chemical weathering with smectite formation and erosion, is recognized around 490-530 cm. However, climate improvement (in terms of precipitation and/or temperature) was insufficient during this interval to allow an extensive precipitation of authigenic calcite in the lake.

5.2. A complex chronological framework

The age model (see discussion in Belcher et al., 2009) (Fig. 4) shows that the Ohrid sequence extends back to about 140,000 yr BP. It fits remarkably well with the orbitally-tuned environmental history of southern Europe (Tzedakis, 2005; Stoebich et al., 2007; Sánchez Goñi, 2007) (Fig. 5) from terrestrial pollen records in Greece (e.g., Tenaghi Phillipon) and marine isotope records along the southern European margin (e.g., ODP 380). It also fits with the chronology of pollen events at Monticchio in Italy (Brauer et al., 2007) (Fig. 6). According to this, the Eemian interglacial started at 127,000 yr BP and lasted roughly 17,000 yr which corroborates earlier estimations in southern Europe (Tzedakis et al., 2002b; Brauer et al., 2007; Brewer et al., 2008).

A major hiatus of roughly 12,000 years corresponds to a large part of the S. Germain interglacial. The stratigraphical and chronological position of this hiatus is confirmed by the absence in our record of sapropel YS, dated at 155,000 ± 2000 yr BP (Keller et al., 1978; Kramm, 1997; Wulf et al., 2006; Diviacco et al., 2005; Paternei et al., 2008). This sapropel has been found in marine sediments from the Ionian Sea (Paternei et al., 2008) and in the lower part of the S. Germain interglacial at Monticchio (Brauer et al., 2007). Dating uncertainties remain for the Melseby 2 interstadial, which is dated to ca 87,000 yr BP at Ohrid and ca 85,000 yr BP at Monticchio. The phase of forest expansion within the last glacial period, which is dated to around 50-45,000 yrs ago at Ohrid, closely corresponds to a warm event recorded in speleothem data from Villars in Southern France (Gerry et al., 2003, 2005). This likely corresponds to the forest phase dated slightly earlier at Monticchio (Brauer et al., 2007), Ioannina (Tzedakis et al., 2002a), and Kopais (Oukla et al., 2001). At Villars, this event has been correlated with Dansgaard-Oeschger event 12 with exceptionally high speleothem growth rate and very low δ^18O values (as low as 1%), attributed to dense vegetation growth. Based on 12 TMS U-series dates, the optimum recorded by 813C values is dated to 45,300 ± 400 cal yr BP. The climate improvement responsible for forest expansion at Ohrid at that time is also reflected in the temporary increase of dissolved calcium concentrations that can be inferred from the discrete occurrence of good ostracode preservation (Fig. 4). It was however not sufficient to significantly revive the formation of authigenic calcite.

5.3. Regional environmental changes

Our multi-proxy record suggests that in response to an improvement of climate conditions, Lake Ohrid switched from a less to more productive ecosystem as early as 135,000 cal yr BP. The predominance of steppic herbaceous pollen types (mainly Artemisia and Chenopodiaceae) before that time indicates that steppe vegetation had dominated
Fig. 4. Lake Core JO2004 multiproxy record showing from left to right: radiocarbon ages, main grainstone units, magnetic susceptibility (on both cores JO2004-1 and JO2004-1x); sediment geochemistry with Sr/Sc (dark blue) ratio and major elements Al₂O₃ (light blue), SiO₂ (purple), Fe₂O₃ (light green) and LOI (brown); bulk mineralogy with clays (purple), quartz (orange), calcite (light blue), and feldspar (black); main clay minerals with illite (dark blue), and smectite (pink); the main pollen types and the sum of arboreal pollen percentages; δ¹³C of Carbondate ostracodes (dark blue). Note the lack of ostracod isotope records during phases of calcite absence.
However, soils were still poorly developed as indicated by the relatively high values of $^{13}$C in ostracod calcite. At 128,000 yr BP, a slight decrease in total arboreal pollen percentages together with an abrupt positive shift in ostracod $\delta^{13}$C of more than 1% and a plateau in calcite content suggest a short episode of climate deterioration interrupting the long-term warming trend. This pattern of environmental change has been recorded in several other climate-proxy records from southern Europe (e.g., Tzedakis et al., 2002a; Brauer et al., 2007; Cossey et al., 2004). An abrupt, simultaneous shift in most proxies at 127,900 yr BP marks the true beginning of the Last Interglacial period. The episode between 126,000 and 122,000 yr BP is characterized by increased calcite precipitation within the lake as recorded by peak CaCO$_3$ content (> 85%) and the large positive excursion of Sr/Th. At that same time, Mediterranean trees expanded at the altitude of Lake Ohrid whereas cold steppes disappeared. Mesic forests developed on the adjacent depresses and conifer forest at higher altitude. This indicates that this period was unambiguously the warmest of the entire last interglacial. According to other data, the thermal maximum was reached around 124,700 yr BP when Mediterranean vegetation elements widely expanding in southern European lowlands (e.g., Rovee et al., 2006) reached the Lake Ohrid area at 800 m altitude. The end of this climatic optimum at 122,000 yr BP is marked by a dramatic shift in forest composition. Temperate and Mediterranean forest trees declined, whereas Cupressus and Abies-dominated mixed forests, then Abies- and Picea-dominated conifer forests widely expanded with Abies proportion higher than those recorded anywhere in southern Europe (e.g., Rovee et al., 2000; Tzedakis et al., 2002a; Pini et al., 2003; Allen and Huntley, 2003), indicating the incursion of mountain environments into the lake basin and a progressive cooling. This is supported by the decreasing trend of calcite and increasing trend of detrital quartz during this interval. However, remarkably stable ostracod $^{13}$C values around $-26\%$ throughout the period 127,000-110,000 yr BP indicate that tree cover and related soil stability in this mountain environment were not affected by climate variations within the Last Interglacial.

Calcite minima and significant expansion of herbaceous steppe plant types at the expense of forest signal significant regional climate deterioration from 110,000 to 107,000 yr BP and around 87,000 yr BP, i.e. coeval with the Mc keys 1 and 2 stadials, respectively. The cold climate conditions during stadials were responsible for poor soil development and enhanced erosion as recorded by the ostracod-isotope and mineralogical records. The Saint Genmain 1 and 2 interstadials are only partly recorded in Core J02004-1 (Fig. 6). Although all proxies record climate improvement around 106,000 and 82,000 yr BP, continued predominance of conifers at the expense of mesic trees in regional forests, and lower calcite content compared to that attained during the Last Interglacial suggest less favourable climate conditions and lower winter temperatures. This differs from what is observed at Ioannina (470 m altitude) in Northern Greece (Tzedakis et al., 2002a) where the two interstadials appear to have been remarkably similar in vegetation and climate conditions.

The abrupt fall in calcite content at the end of Saint Genmain 2 (around 79,000 yr BP) shows that a lower temperature threshold was crossed, leading to at least seasonally frozen soil conditions at the elevation of the lake and probably permafrost in a large part of the mountain system. This is supported by the occurrence of periglacial landforms observed near the top of the Galicica mountain range (Belnocheri et al., 2009). Cold-tolerant trees such as Juniperus and Betula expanded together with steppe herbaceous plant types similar to what occurred widely in southern Europe during the last glacial period (e.g., Allen et al., 1999; Allen and Huntley, 2003). However, the continuous presence of deciduous trees (albeit in low percentages) suggests that the forest refuge for mesic trees present at lower elevation at Ioannina (Tzedakis et al., 2002a) and in the nearby Mavro-Basin (Denielic et al., 2000) expanded up to 700-800 m in altitude to the shores of Lake Ohrid. The moderating lake effect on regional climatic conditions probably favoured their development at middle

![Fig. 5. Comparison between two independent chronologies of environmental changes in southern Europe. Upper panel (A) June insololation curves for 67°N. (B) benthic foraminalinal $\delta^{18}$O from ODP 989 (latitude: 55°29'11" N, longitude: 14°42'10" W), representing the ice volume/seas-level component of north-south $\delta^{18}$O from core ODP 989, and (C) the tree-pollen percentage curve from Thraci Phippedou plotted on the time scale of Tzedakis et al. (2003) (redrawn from Tzedakis (2005); see references therein). Lower panel: June insololation curves for 67°N (Reeve, 1978), pollen percentage curves for steppe elements and tree, and the percent calcite content of core J0-2004 1 (this study) plotted on the same time scale. Grey vertical bars show the correspondence between the two independent data sets.](image-url)
altitudes in southern Albania. High-amplitude variation in the mineralogical composition of Ohrid sediments deposited between 50,000 and 18,000 cal yr BP indicate unstable environmental conditions during the glacial period. For example, alternating maxima of illite and smectite in the clay-mineral assemblages suggest variable erosion and weathering rates due to fluctuation of precipitation and/or temperature, with illite maxima corresponding to intervals of enhanced precipitation and erosion.

The presence of refugia for mesic trees originating from the forest phase developed around 50–45,000 yr BP, likely explains the surprisingly progressive last glacial-interglacial transition (Termination I) observed at Ohrid and Kastoria (Zredakis, 2005) with mesic tree pollen types increasing regularly from 38,000 BP to the Holocene (8000 cal yr BP). Authigenic calcite precipitation started to increase only at 17,000 cal yr BP, illustrating the delayed response of the lacustrine system to late-glacial warming. The increase of calcite was interrupted during the Younger Dryas, and then restarted from ca. 11,000 cal yr BP to reach a maximum at 6000 cal yr BP. This rising trend was interrupted by an abrupt event of forest degradation probably coeval with the “8.2 kyr” cold event already recorded at Maliq (Bordon et al., 2006), corresponding to a −2 °C cooling of annual temperature compared to mean Holocene values.

6. Concluding remarks

The most prominent feature in the recorded history of Lake Ohrid is the contrast between the responses of lacustrine and terrestrial environments to climate change throughout the last climatic cycle. Abrupt hydrological changes were controlled by the karstic system, which was active during interglacials and completely blocked during glacials in response to drastic deterioration of climate conditions. The abundance and strontium content of authigenic calcite and the δ13C of ostracod calcite appear to be sensitive proxies of temperature change during interglacials and help to detect and characterize the main fluctuations of climate within the interglacials and at the glacial-interglacial transitions. In contrast, the terrestrial environment recorded a more gradual evolution, particularly at the last glacial-interglacial transition. The Riss–Eemian (Termination II) and Würm–Holocene (Termination I) climate transitions significantly differ from one another, with the lacustrine system reacting well before (by about 10,000 years) local vegetation change at the Riss–Eemian transition, and peak lake productivity lagging forest expansion by several thousand years at the Würm–Holocene transition. The interruption of the Würm glaciation by a more moderate forest phase around 50–45 kyr, is probably one of the main causes of the presence of forest refugia.
which enabled, together with probable moderating effect of a large lake on local climate, the early expansion of forests during the Würm-Holocene transition. We suggest that during the Riss glaciation no such moderate forest phase existed. The Riss glaciation is well known to have developed larger glaciated areas around the Alps than the Würm (van Halen, 2004) and there is evidence of large, pneumal deposits in the area of Lake Ohrid. However, the corresponding moraines are not yet dated and continuous pollen records for the Riss glaciation are still lacking, in contrast, the delayed lake response to global warming at the last glacial-interglacial transition, shown by calotte production peaking at 50,600 cal yr BP, illustrate the threshold response of the geosystem which was completely blocked during the late Würm glaciation. The succession of abrupt climate changes — especially the Bolling-Allerød Younger Dryas sequence and the "8.2kyr cold event" — in this mountainous area probably accentuated this delay.

By using a large range of paleoenvironmental proxies we were able to detect climate changes even during glacial phases when calcite and the associated stable isotopic records are lacking. This study illustrates the great potential of Lake Ohrid as an archive of southern European climate and environment over the entire Quaternary.

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