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ORIGINAL PAPER

# A paleoclimate record with tephrochronological age control for the last glacial-interglacial cycle from Lake Ohrid, Albania and Macedonia

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**Abstract** Lake Ohrid is probably of Pliocene age, and the oldest extant lake in Europe. In this study climatic and environmental changes during the last glacial-interglacial cycle are reconstructed using lithological, sedimentological, geochemical and physical proxy analysis of a 15-m-long sediment succession from Lake Ohrid. A chronological framework is derived from tephrochronology and radiocarbon dating, which yields a basal age of ca. 136 ka. The succession is not continuous, however, with a hiatus between ca. 97.6 and 81.7 ka. Sediment accumulation in course of the last climatic cycle is controlled by the complex interaction of a variety of climate-controlled parameters and their impact on catchment dynamics, limnology, and hydrology of the lake. Warm interglacial and cold glacial climate conditions can be clearly distinguished from organic matter, calcite, clastic detritus and lithostratigraphic data. During interglacial periods, short-term fluctuations are recorded by abrupt variations in organic matter and calcite content, indicating climatically-induced changes in lake productivity and hydrology. During glacial periods, high variability in the contents of coarse silt to fine sand sized clastic matter is probably a function of climatically-induced changes in catchment dynamics and wind activity. In some instances tephra layers provide potential stratigraphic markers for short-lived climate perturbations. Given their widespread distribution in sites across the region, tephra analysis has the potential to provide insight into variation in the impact of climate and environmental change across the Mediterranean.

**Keywords** Lake Ohrid · Mediterranean · Tephrochronology · Paleolimnology · Last glacial-interglacial cycle

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## Introduction

Large ancient lakes have been widely recognised as extremely important continental archives of long-term climatic and environmental changes (Colman et al. 1995; Zolitschka et al. 2006; Melles et al. 2007). Lake Ohrid is located within an active subsidence zone in the northeastern Mediterranean (Fig. 1) and is probably between five and two million years old (Stankovic 1960; Albrecht and Wilke 2008).

Most continental paleoclimate records from the Mediterranean region are restricted to the late last glacial or the Holocene and only few cover longer periods (Bar-Matthews et al. 2000; Martrat et al. 2004; Brauer et al. 2007; Tzedakis et al. 2003). Given its age and geographical position at the interface of the large-scale atmospheric Hadley and Mid-Latitude cells, Lake Ohrid fulfils an important set of prerequisites necessary to gain a better understanding of Mediterranean climate evolution over longer time scales.

Precise and accurate dating is a major precondition for a robust interpretation and comparison of paleoclimate records with those from the surrounding. In the Mediterranean region tephra layers, can provide independent stratigraphical tie points beyond the limit of radiocarbon dating, since intense and perpetual volcanic activity has led to the widespread dispersal and deposition of tephra layers across the region during the Quaternary (Giaccio et al. 2008). Previous tephrostratigraphical studies of lake-sediment records from Lake Ohrid have revealed that it represents an excellent tephrostratigraphic archive (Wagner et al. 2008a; Vogel et al. 2009).

Studies focussing on the recent response of Lake Ohrid to climatic and environmental change and anthropogenic impact have shown that a range of limnological parameters are sensitive to subtle changes in temperature and nutrient supply (Matzinger et al. 2006a, 2007). So far four longer records have been recovered from different parts of the lake (Roelofs and Kilham 1983; Wagner et al. 2009; Belmecheri et al.

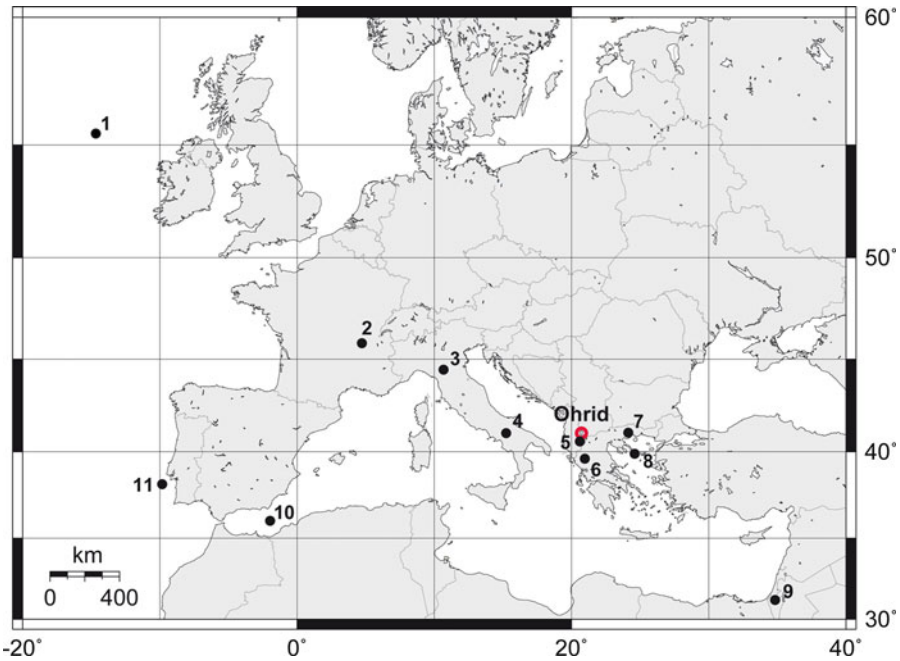


**Fig. 1** Map showing the location of Lake Ohrid (*red circle*) on the Balkan Peninsula in the central/northern Mediterranean and sites referred to in the text. 1. ODP site 980 (McManus et al. 1999), 2. Paleo lake Les Echets (Wohlfarth et al. 2008), 3. Corchia cave (Drysdale et al. 2005; Zanchetta et al. 2007), 4. Lago Grande di Monticchio (Allen et al. 1999; 2002; Brauer et al. 2007; Allen and Huntley 2009), 5. Lake Maliq (Bordon et al. 2008), 6. Ioannina (Tzedakis et al. 2003, Lawson et al. 2004), 7. Tenaghi Phillippon (Kotthoff et al. 2008), 8. SL 152 (Kotthoff et al. 2008), 9. Soreq cave (Bar-Matthews et al. 1999, 2000), 10.ODP site 977A (Martrat et al. 2004), 11. MD95-2042 (Sánchez Goñi et al. 2000)

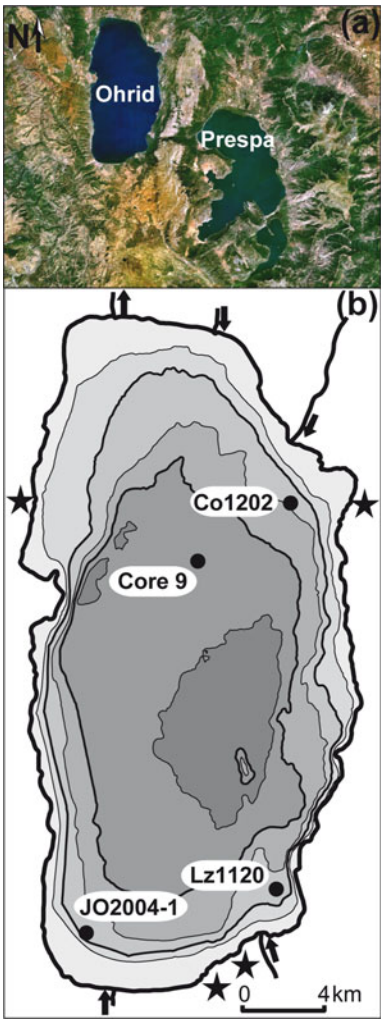


**Fig. 2** **a** Satellite image showing Lakes Ohrid and Prespa. **b** Bathymetric map of Lake Ohrid with 50 m isolines showing the location of coring sites Co1202 (Vogel et al. 2009), Core 9 (Roelofs and Kilham 1983), Lz1120 (Wagner et al. 2008a, 2009), and JO2004-1 (Belmecheri et al. 2009). Main riverine inflows are indicated by *arrows* pointing towards the lake and the outlet is indicated by the *arrow* pointing away from it. Major subaerial spring areas are indicated by *asterisks*

2009; Vogel et al. 2009; Fig. 2). Together, these records indicate marked limnological contrasts between glacial and interglacial phases at Lake Ohrid and confirm its potential for longer-term paleoclimate analysis. However, detailed analysis of limnological and hydrological variability and changes in catchment dynamics is still lacking for the full last glacial-interglacial cycle, and the glacial-interglacial transition in particular.

In this study lithological, sedimentological, geochemical, and physical properties are used to assess climatic and environmental changes during the last glacial-interglacial cycle and back to MIS 6, based on an approximately 15-m-long sediment record (core Co1202) from the northeastern part of Lake Ohrid. The chronological framework of the Co1202 record is primarily based upon tephrochronology (Vogel et al. 2009). Although core Co1202 has an erosional hiatus between ca. 96 and 81 kyears BP (Vogel et al. 2009) it seems otherwise to be complete and hence allows a comparison of its deduced climatic and environmental changes with other records from Lake Ohrid as well as the wider Mediterranean region.

## Site description

Lake Ohrid (41°01′N, 20°43′E; Figs. 1 and 2), a transboundary lake shared by the Republics of Albania and Macedonia, is situated at 693 m above sea level (a.s.l.) and surrounded by high mountain ranges of up to 2300 m a.s.l. The lake is approximately 30 km long, 15 km wide and covers an area of 358 km$^2$. The lake basin has a relatively simple tub-shaped morphology with a maximum water depth of 289 m (Fig. 2), an average water depth of 151 m, and a total volume of 50.7 km$^3$ (Popovska and Bonacci 2007).

The topographic watershed of Lake Ohrid comprises an area of 2393 km$^2$ incorporating Lake Prespa, which is situated 10 km to the east of Lake Ohrid at an altitude of 848 m a.s.l. (Popovska and Bonacci 2007). The two lakes are connected via karst aquifers passing through the Galicica and Suva Gora mountain range. Karst springs depleted in nutrients and minerogenic load represent the primary hydrologic inputs to Ohrid (55%). Up to 50% of the water emerging from the inflow springs originates from Lake Prespa, with the remainder coming from precipitation on the surrounding mountain ranges (Anovski et al. 1992; Matzinger et al. 2007). Direct precipitation on the lake surface, river- and direct surface runoff together account for the remaining 45% of the hydrologic input to Lake Ohrid. Surface outflow (60%) through the river Crn Drim to the north and evaporation (40%) are the main hydrologic outputs (Matzinger et al. 2006a). Due to the overall low nutrient load of the hydrologic inputs, inhibited

resuspension of nutrients from the sediment–water interface by incomplete and irregular mixing of the water body, and its large volume, Lake Ohrid is highly oligotrophic (Stankovic 1960; Matzinger et al. 2007).

Due to its sheltered position in a relatively deep valley surrounded by high mountain ranges and to its proximity to the Adriatic Sea, the climate of the Lake Ohrid watershed shows both Mediterranean and continental influences (Watzin et al. 2002), with an average annual air temperature for the period between 1961 and 1990 of 11.1°C, maximum temperatures below 31.5°C, minimum temperatures above −5.7°C, and an average annual precipitation of 800–900 mm (Popovska and Bonacci 2007). Prevailing wind directions follow the N–S axis of the Ohrid valley.

The Ohrid valley is located within an active tectonic graben system of the Western Macedonian geotectonic zone, which is part of the interior Dinaric Alps. Metamorphic and magmatic rocks of Paleozoic age are exposed to the north and northeast of the valley. Strongly karstified limestones and clastic sedimentary rocks of Triassic age are exposed along the northwestern, eastern, and southeastern shorelines (Watzin et al. 2002). Ultramafic rocks and associated weathering crusts containing chromium and iron-nickel ore deposits of Jurassic-Cretaceous age outcrop along the southwestern and western shorelines (Dilek et al. 2007). Quaternary lacustrine and fluvial deposits occupy the plains to the north and to the south of the lake (Watzin et al. 2002).

## Materials and methods

### Core recovery

Core Co1202 was recovered in autumn 2007 from the northeastern part of the lake basin, where a hydroacoustic survey indicated a water depth of 145 m (Fig. 2) and a widely undisturbed sediment succession with planar bedding and parallel reflectors. The surface sediments and deeper sediments were collected using a 0.6-m gravity corer and a 3-m-long percussion piston corer, respectively (both UWITEC Co.). The overlapping 3-m-long core segments were subdivided into 1-m-long segments in the field. In the laboratory the 1-m-long core segments were split in two halves, wrapped and stored at 4°C.

### Analytical work

One of the core halves was used for high-resolution X-ray fluorescence (XRF) and magnetic susceptibility (MS) measurements. XRF-analysis was carried out using an ITRAX core scanner (COX Ltd.), equipped with a Mo-tube set to 30 kV and 30 mA and a Si-drift chamber detector. Scanning was performed at 1-mm resolution and an analysis time of 20 s per measurement. Data were smoothed using a 5-pt running mean to reduce noise. The obtained count rates for Ti, Zr, Cr, Fe, and Mn can be used as estimates of the relative concentrations for these elements (Croudace et al. 2006). MS measurements were performed at 1-mm resolution using a Bartington system equipped with a high-resolution MS2E sensor.

The other core halve was subsampled continuously at 2-cm intervals. Aliquots of the freeze-dried subsamples were ground to a particle size below 63 µm using a planetary mill for subsequent biogeochemical analyses. Concentrations of total carbon (TC), total nitrogen (TN), and total sulphur (TS) were measured with a Vario Micro Cube combustion CNS elemental analyzer (VARIO Co.). Samples for total organic carbon (TOC) analysis were pre-treated with HCl (10%) at a temperature of 80°C to remove carbonates and then analyzed using a Leco CS-225 carbon–sulfur detector (LECO Corp.). The amount of total inorganic carbon (TIC) was determined from the difference between TC and TOC. The calcite ($CaCO_3$) content was calculated from TIC under the assumption that TIC solely originates from $CaCO_3$. Biogenic silica (BSi) concentrations were assessed using Fourier transform infrared spectroscopy (FTIRS) and a partial least square calibration model build upon 119 samples of core Lz1120 from the southeastern part of the lake (Vogel et al. 2008). For a detailed description of the method and sample pretreatment prior to analysis see Vogel et al. (2008) and Rosén et al. (2009).

Grain-size analyses were carried out on clastic detritus. For this purpose $CaCO_3$, finely dispersed iron sulphides, organic matter, and BSi were removed from the subsamples using standard methods. Grain-size distribution was measured using a Micromeritics Saturn Digisizer 5200 laser particle analyzer. Grain-size distributions are calculated from the average values of 3 runs and given as volume percentages (vol%) of the individual grain-size fractions.

The chronology of core Co1202 is discussed in detail in Vogel et al. (2009) and based on ten encountered tephra and cryptotephra layers and on radiocarbon dating of bulk organic matter carbon on seven samples. According to Vogel et al. (2009) radiocarbon ages obtained on the humic acid (HA) fraction yielded the most reliable ages and were calibrated into calendar years before present (cal yrs BP) using CalPal-2007$^{online}$ and the CalPal2007_HULU calibration curve (Danzeglocke et al. 2008).

## Results and discussion

### Lithology and Lithofacies classification

Individual core segments were correlated using magnetic susceptibility (MS), x-ray fluorescence (XRF) data, and lithological core descriptions, thus leading to a composite core of 14.94-m length. The succession comprises two distinctly different lithofacies, which can further be subdivided into sublithofacies (1a, b, c & 2a, b; Fig. 3).

Lithofacies 1 (1494–1439, 1064–246 cm; Fig. 3) appears light-grey to dark-grey and black and is mainly composed of clastic clayey-sandy silts. Frequent occurrence of coarse sand and gravel grains is interpreted as ice rafted detritus (IRD). Peaks in clastic material correlate well with high Ti and MS values. Tephra and cryptotephra layers at 1,447–1,440, 825–822, 752–743, 696–689, 620–617, and 277.5–269 cm (Vogel et al. 2009) explain maxima in MS values at 746, 693, 619, and 275 cm and minima in Ti intensities at 746, 693, and 619 cm (Fig. 3). Carbonates are almost absent in Lithofacies 1, apart from the section between 362 and 264 cm. The amount of finely-dispersed organic matter, as also indicated by the TOC content, is relatively low throughout Lithofacies 1 but shows some subtle variation. Terrestrial and subaquatic plant macrofossils are absent. Poorly preserved diatom frustules occur in relatively low abundance as also indicated by low BSi concentrations. Based on more subtle changes in sediment structure and composition Lithofacies 1 can further be divided into three sub-lithofacies (1a, b, c; Fig. 3). Lithofacies 1a (1494–1439, 949–825, 727–634 cm) appears dark-grey to black, massive, with clear signs of bioturbation and an intermediate content of finely dispersed OM compared to other sublithofacies. Lithofacies 1b (1064–949, 825–727, 634–362, 264–246 cm) appears light grey, with irregular laminations, concretionary horizons, and the lowest content of OM within Lithofacies 1. Laminations thicknesses varies from 0.5 to 2 mm, are dark grey to dark green in colour, and often occur as stacked sequences composed of several individual layers. Concretionary horizons show thicknesses of up to 2 cm, are dark grey and often contain vivianite crystals. Lithofacies 1c (362–264 cm) appears light grey, with irregular laminations and the highest OM contents within Lithofacies 1; it contains finely dispersed carbonates.

Lithofacies 2 (1439–1097, 246–0 cm; Fig. 3) appears either light-brown to light-grey or dark-brown to black and contains significant amounts of clay to fine-silt sized carbonates and generally low but varying amounts of clay to fine-sand sized detrital clastic material. Coarse sand and gravel is absent throughout Lithofacies 2. The high carbonate content and associated low content of detrital clastic material correlates well with the low Ti and MS values. Peak MS values centred around 1229, 1143, 145, 76 cm correspond to the occurrence of tephras and cryptotephras (Vogel et al. 2009). Well-preserved carbonate microfossils (ostracods) are abundant, while terrestrial and subaquatic plant macrofossils are absent. The content of OM is higher than Lithofacies 1, as is the abundance and preservation quality of diatom frustules. A subdivision of Lithofacies 2 into two sub-lithofacies is based on colour variations and changes in carbonate and clastic matter content. Lithofacies 2a (1439–1402, 1271–1209, 1122–1095.5, 246–222, 106–52, 25–0 cm) appears dark-brown to black, with lower carbonate contents and higher content of clastic detritus. Lithofacies 2b (1402–1271, 1209–1122, 222–106, 52–25 cm) appears light brown to light-grey, with the maximum carbonate and minimum detrital clastic content within the sediment sequence as a whole. Both SEM screening and BSi values indicate better diatom preservation in sub-lithofacies 2b than sub-lithofacies 2a.

At 1095.5–1064 cm Lithofacies 1 and 2 are separated by a graded sandy horizon with an apparent erosive base (Fig. 3), such as indicative of a mass-wasting episode. Mass wasting due to earthquakes or lake level changes is common especially in the lateral parts of the lake basin (Wagner et al. 2008b) and also might have caused the erosional hiatus in core

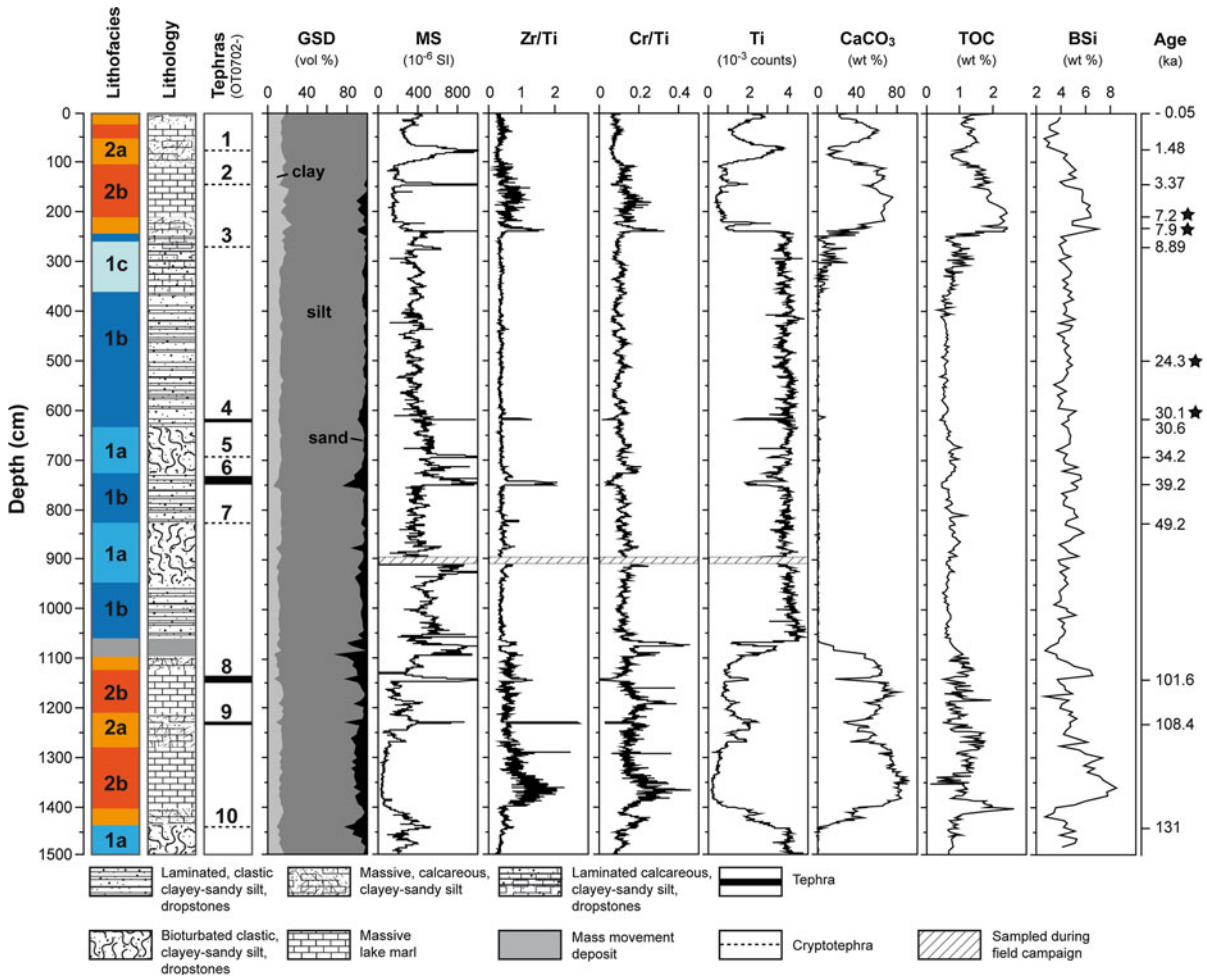


**Fig. 3** Lithofacies, lithology, position of tephra and cryptotephra layers, grain-size distribution (GSD), magnetic susceptibility (MS), Zr/Ti, Cr/Ti ratios, Ti intensities, calcite ($CaCO_3$), TOC, and biogenic silica (BSi) contents of core Co1202. Ages to the right of the figure are derived from correlations of tephra and cryptotephra horizons and radiocarbon dating (*asterisks*)

Co1202, which was recovered from a slightly inclined slope approximately 2.5 km offshore.

## Chronology of core Co1202

For the establishment of an age-depth model based on the tephrostratigraphic time markers and the radiocarbon dates (Fig. 3), the year of the coring campaign (2007) was taken as the date of the sediment–water interface. After removal of event layers, such as discrete tephra layers at 1232.5–1229 cm (OT0702-9), 1146.5–1140 cm (OT0702-8), 752–743 cm (OT0702-6), 620–617 cm (OT0702-4), and the mass-wasting deposit at 1095.5–1064 cm (Fig. 3), chronological tie points were interpolated linearly, including a proposed a hiatus of approximately 16 ka between 97,6 and 81,7 ka (Fig. 3). The basal age for core Co1202 was extrapolated to ca. 136 ka (Vogel et al. 2009).

## Indicators for environmental and climatic change

### *TOC content*

The TOC content represents the amount of finely dispersed OM and tracks the subdivision of core Co1202 into lithofacies and sub-lithofacies (Fig. 3). OM found in Lake Ohrid sediments is mainly of autochthonous origin with only minor contribution from allochthonous sources (Matzinger et al. 2007; Wagner et al. 2008b, 2009) and is thus strongly

controlled by primary productivity in the lake. Water temperature, light availability, nutrient supply from river inflows and to a minor amount from karst springs, and nutrient replenishment from the sediment–water interface are the principal factors controlling productivity in Lake Ohrid today (Matzinger et al. 2006a, b, 2007).

OM concentration will also be influenced by rates of decomposition. Annual mixis of Lake Ohrid′s water column and high oxygen saturation down to 150 m water depth and site Co1202, such as observed today (Matzinger et al. 2006a), supports the decomposition of OM.

*Calcite content ($CaCO_3$)*

The majority of $CaCO_3$ in Lake Ohrid sediments can be attributed to photosynthesis-induced inorganic calcite precipitation (Wagner et al. 2008b, 2009; Belmecheri et al. 2009), with only minor contribution from biogenic sources such as ostracod valves (Belmecheri et al. 2009; F. Viehberg pers, commun,).

Calcite precipitation in Lake Ohrid is supposed to mainly occur during spring-summer in the epilimnion, when photoautotrophic organisms assimilate $CO_2$ (Wetzel 2001; Matzinger et al. 2007) and as long as supply of $Ca^{2+}$ and $HCO_3^-$ ions is maintained. $Ca^{2+}$ and $HCO_3^-$ ions are mainly delivered to Lake Ohrid by karstic inflows (Matzinger et al. 2006b), whilst surface runoff and direct precipitation dilute the water of Lake Ohrid with respect to $Ca^{2+}$ and $HCO_3^-$ ion concentrations (Belmecheri et al. 2009). The concentration of $Ca^{2+}$ and $HCO_3^-$ ions is also modified by the rate of evaporation, which is primarily controlled by temperature and wind intensity.

Dissolution of calcite in the water column and in surface sediments is primarily triggered by aerobic decomposition of OM, which produces significantly more carbonic acid ($H_2CO_3$) than anaerobic decomposition of OM (Müller et al. 2006). Hence, relatively high availability and replenishment of $O_2$ in the water column as observed at the coring location today would support dissolution of calcite, as long as decomposition of OM takes place.

*Grain-size distribution (GSD)*

The grain-size distribution of clastic detritus in core Co1202 serves as an indicator for transport processes and energies. Fine to medium-sized silts with lower but variable contents of coarse silt comprise 60–85 vol% of the clastic fraction. Fluctuations in the proportion of clay (5–23 vol%) and sand (0–37%) drive the main variations in GSD throughout core Co1202 (Fig. 3). Coarse sand and gravel have not been quantitatively assessed, but are restricted to Lithofacies 1 sediments. Most of the sand-sized clastics are fine sands (0–35 vol%) with only minor contribution from medium sands (<3 vol%). Peaks in sand content at 1438, 1229, 1140, 752, and 617 cm correspond to tephra and/or cryptotephra layers, while those between 1095.5 and 1064 cm are related to a mass movement deposit (Fig. 3; Vogel et al. 2009).

Clay and fine to medium silt-sized clastics in lakes can easily be transported over long distances by wind-induced or riverine currents. The transport of fine to medium-sized sand requires higher transport energy. Since site Co1202 is situated 2.5 km from the nearest shore and is not in close proximity to any major river inflow (Fig. 2) it can be assumed that sedimentation of fine to medium-sized sand at site Co1202 requires significantly stronger wind-induced surface currents or ice floe transport.

*Titanium (Ti)*

Ti can occur in a wide range of relatively insoluble mineral phases and is therefore commonly applied as indicator for aeolian (Yancheva et al. 2007) or fluvial clastic input in lacustrine systems (Minyuk et al. 2007). Since the clastic detritus in core Co1202 is dominated by fine to medium-sized silts, and there is a negative correlation between the sand content and Ti intensities (Fig. 3) it can be speculated that Ti intensity in Lake Ohrid sediments is principally associated with fine-grained, clay to silt-sized clastic material and thus mainly derived from fluvial input.

*Zirconium (Zr) and Zr/Ti ratio*

Zr is regarded as a relatively immobile element primarily originating from the mineral zircon, which has a high density (4.7 g/cm$^3$) and resistance to chemical and physical weathering; it is therefore probably transported along with coarser grain-size fractions. To account for dilution mainly by authigenic calcite in Lithofacies 2 and 1c, the Zr/Ti ratio is

used. Highest Zr/Ti ratios generally occur in Lithofacies 2b and in particular between 1390 and 1330 cm. Despite this general pattern peak Zr/Ti ratios at 1230, 1144, 825, 746, and 618 cm are related to tephra or cryptotephra layers (Fig. 3; Vogel et al. 2009). Lower but highly variable Zr/Ti ratios are documented for Lithofacies 1. In previous studies Zr contents have been applied as a proxy for wind intensities and aeolian clastic material (Müller et al. 2001). In the case of Lake Ohrid the Zr/Ti ratio serves as a high-resolution indicator of coarser silt to sand-sized clastic matter and thus as a proxy for the intensity of wind-induced surface currents and erosion in the catchment.

*Chromium (Cr) and Cr/Ti ratio*

Cr in sediments of Lake Ohrid probably derives from Cr ore bearing ultramafics and associated weathering profiles situated along its south-western and western shore line (Albanian geological survey 1999; Spirkovski et al. 2001). The Cr/Ti ratio is used to account for dilution by authigenic calcite (Fig. 3) and thus defines variations of the Cr content within the clastic fraction of the sediments. Highest Cr/Ti ratios generally occur in Lithofacies 2b and in particular between 1390 and 1330 cm. Lower but strongly variable Cr/Ti ratios are documented in Lithofacies 1. Cr is likely transported to site Co1202 by a wind-driven, counter-clockwise surface current (Stankovic 1960) and thus serves as an indicator for wind intensities and erosion rather than different redox conditions or variations in OM content, which are also known to influence Cr concentrations in lake sediment (Schaller et al. 1997; Cohen 2003).

*Concretionary Fe/Mn horizons (Fe/Mn redox fronts)*

Clearly visible concretionary horizons occurring in Lithofacies 1b (Fig. 3) correspond well to peak Fe and Mn intensities in core Co1202 (ESM1). Similar horizons enriched in Fe and Mn have previously been described in Lake Baikal (Granina et al. 2004; Fagel et al. 2005). At sites with low sedimentation rates in Lake Baikal Fe/Mn horizons today form at redox fronts 15 to 25 cm below the sediment–water interface (Granina et al. 2004; Fagel et al. 2005), hence the depth of oxygen penetration generally marks the upper boundary of the redox front (Wetzel 2001). Slow dissolution and reprecipitation causes a continuous upward movement of these concretionary horizons as long as the sedimentation regime remains constant (Granina et al. 2004). Changes in the sedimentation regime can lead to burial and preservation of these Fe/Mn horizons (Deike et al. 1997; Granina et al. 2004; Demory et al. 2005). Therefore, Fe/Mn concretionary horizons found in Lithofacies 1b sediments in core Co1202 probably represent buried paleo redox fronts and can be used as indicators for low sedimentation rates, well-oxygenated water and upper sediment column, and rapid and recurrent changes in sedimentation regime.

## Discussion

### The penultimate glacial and termination II (135.9–127.3 ka)

Lithofacies 1a sediments (1494–1439 cm; 135.9–130.3 ka) represent environmental and climatic conditions of MIS 6 in core Co1202. Marginal variations in sediment composition, which is dominated by clastic detritus and only low amounts of biogenic and authigenic material (Fig. 4) imply a relatively stable low productivity.

Occurrence of gravel-sized clastic detritus requires transport by ice floes. This implies that Lake Ohrid was at least partly ice covered during winter. Given that Lake Ohrid today only rarely freezes (Stankovic 1960) it can be assumed that winter temperatures were significantly lower towards the end of MIS 6, which is well supported by pollen-based temperature reconstructions in the region (Allen and Huntley 2009).

The stable low productivity environment is probably a result of persistent cold climate conditions in the Mediterranean (Martrat et al. 2004; Allen and Huntley 2009) and further implies that surface water temperatures during spring and summer were significantly lower. The supply of nutrients by surface runoff, rivers and via karst aquifers from Lake Prespa is difficult to infer for the penultimate glacial. It can, however, be speculated that nutrient supply by surface runoff and rivers was reduced due to a relatively sparse steppe-like vegetation cover in the region (Tzedakis et al. 2003; Allen and Huntley 2009), which resulted in poorly developed soils at Lake Ohrid. Complete mixis

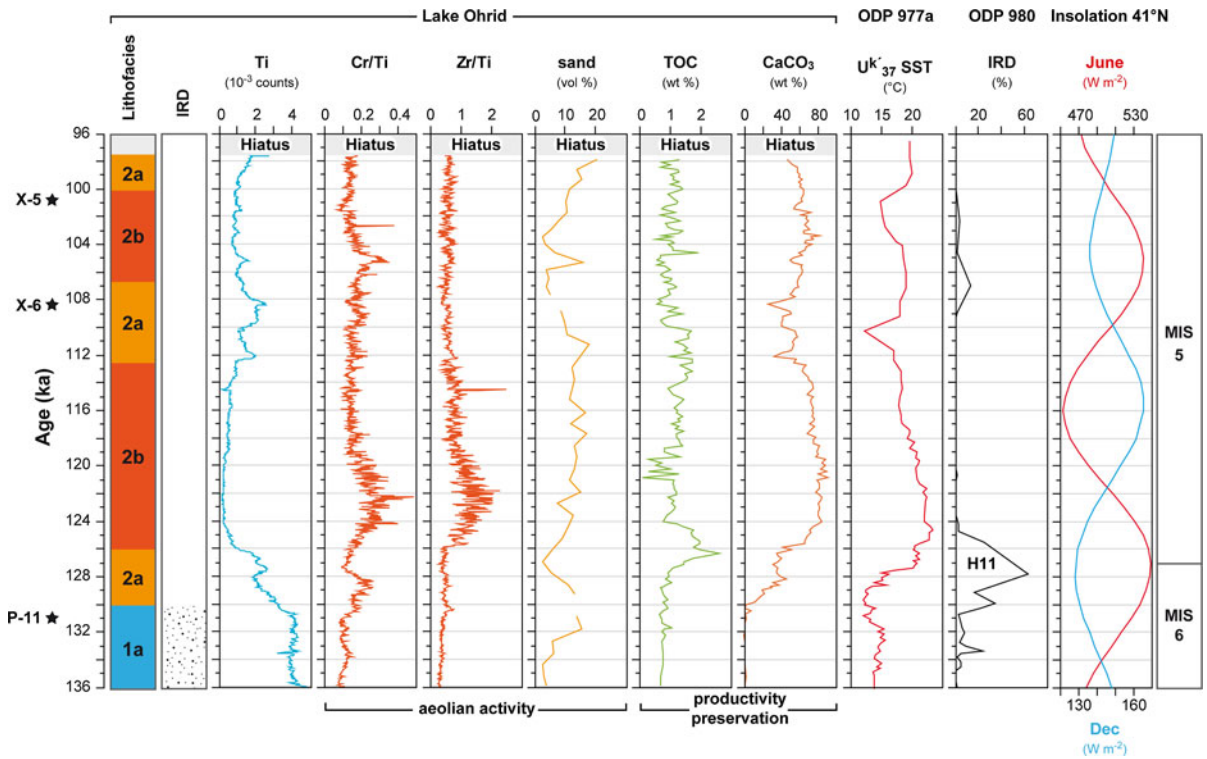


**Fig. 4** Lithofacies (legend shown in Fig. 3), IRD occurrences, Ti intensities, Cr/Ti, Zr/Ti ratios, sand, TOC, $CaCO_3$ contents, and tephrochronological age control points (*asterisks*) of core Co1202 between ca. 136 and 97.6 ka. Also shown are $U^{K'}_{37}$ derived sea surface temperatures from ODP site 977a (Martrat et al. 2004), IRD contents from ODP site 980 (McManus et al. 1999), the local June and December (Dec) insolation (Berger and Loutre 1991), and marine isotope stages (MIS; Bassinot et al. 1994). H11 = Heinrich event 11

once or twice a year promoted by low winter temperatures would have fostered aerobic decomposition of OM and dissolution of calcite at the sediment surface (Müller et al. 2006) and restricted recycling of nutrients (Wetzel 2001; Wagner et al. 2009). Low calcite contents in core Co1202 during MIS 6 are probably supported by low $Ca^{2+}$ and $HCO_3^-$ ion concentrations. Sparse vegetation coverage and poorly developed soils hampered the generation of humic and carbonic acids from organic litter, a necessity for the dissolution of the calcareous bedrock. Local ice caps and frozen surfaces on the surrounding mountain ranges (Belmecheri et al. 2009) probably reduced the amount of water percolating through crevasses and promoted surface runoff and erosion, particularly due to pulsed spring-summer melt water discharges. Additionally, lower temperatures and a semipermanent ice cover restricted evaporation. Hence, Lake Ohrid was primarily fed by surface runoff and direct precipitation with only minor contribution of karst springs between 135.9 and 130.3 ka. This together with restricted productivity explains the relatively high amounts of clastic detritus.

An increase in the amount of fine sand and concomitant increases of Cr/Ti and Zr/Ti ratios between 134 and 127.8 ka, with maximum values at 129 ka (Fig. 4), implies stronger wind-induced surface currents and/or enhanced erosion of dry and sparsely vegetated soils in the catchment of Lake Ohrid. The timing may correspond to the interruption of North Atlantic deep-water formation associated with Heinrich Event 11 (McManus et al. 1999; Fig. 4). These Heinrich events are known to promote a cold and dry climate in the Mediterranean region (Sánchez Goñi et al. 2000; Wohlfarth et al. 2008).

The transition from a stable low productivity to a more productive and unstable environment is confined within a relatively short period and represented by Lithofacies 2a sediments between 1439 and 1402 cm (Figs. 3, 4; 130.3–126.3 ka). Rising calcite content

and the absence of ice-rafted detritus (IRD) from 130 ka (Fig. 4) are indicators for the transition from cold glacial to warmer interstadial climate conditions at Lake Ohrid and seem to be roughly synchronous with other paleoclimate records in the Mediterranean (Tzedakis et al. 2003; Martrat et al. 2004; Drysdale et al. 2005; Brauer et al. 2007; Allen and Huntley 2009; Fig. 4). After a plateau-like transitional phase, which lasted until 127.3 ka, a rapid increase in OM content contemporaneous with a stepwise increase in calcite contents probably marks the onset of stronger productivity and Last Interglacial (LI) climate conditions at Lake Ohrid (Fig. 4).

## MIS 5 (127.3–97.6 ka)

Sedimentation under environmental and climatic conditions of MIS 5.5, 5.4, and 5.3 between 127.3 and 97.6 ka is characterised by Lithofacies 2 sediments (1402–1095.5 cm) with absence of IRD, generally high but fluctuating contents of authigenic calcite and OM, and low but fluctuating amounts of clastic detritus (Figs. 3, 4). The distinct variability of these proxies and in grain-size distribution during MIS 5 suggests rather unstable climatic and environmental conditions.

The lack of IRD implies that winter temperatures were high enough to inhibit ice-cover formation at Lake Ohrid. High contents of OM and in particular calcite (Fig. 4) are likely the result of increased spring-summer temperatures and/or reduced mixis of the water column, which are triggering productivity and decomposition. High calcite contents were probably supported by hydrological changes in the catchment. A denser vegetation cover in the region (Tzedakis et al. 2003) probably led to a decrease in surface runoff and erosion and promoted the formation of well-developed soils in the catchment with a stronger production of humic and carbonic acids. In combination with stronger evaporation and increased supply of spring water the $Ca^{2+}$ and $HCO_3^-$ ion concentrations were relatively high during MIS 5. Overall, the calcite content during MIS 5 at Lake Ohrid seems to primarily reflect changes in temperature driven by direct and indirect mechanisms. Maximum calcite values between 124 and 120 ka likely represent maximum temperatures during the LI, which coincides closely with temperature reconstructions from the Western Mediterranean (Martrat et al. 2004) and hydrological changes reconstructed from speleothems from central Italy (Drysdale et al. 2005), and Israel (Bar-Matthews et al. 2000). This period is furthermore characterised by high fine sand contents and maximum Cr/Ti and Zr/Ti values indicating strong wind induced surface currents at Lake Ohrid.

According to the gradually decreasing calcite content, temperatures probably decreased after 120 ka at Lake Ohrid (Fig. 4). This pattern matches well with the gradual temperature decrease reconstructed in the Western Mediterranean (Martrat et al. 2004; Fig. 4), at Lago Grande di Monticchio (Allen and Huntley 2009) as well as a phase of forest opening at Ioannina and Tenaghi Philippon (Tzedakis et al. 2003). Following this rather gradual cooling trend, a more rapid, stepwise decrease in calcite contents from 114.4 ka to 112 ka probably marks the end of LI climate conditions at Lake Ohrid.

Following a period of warmer climate conditions between 111.6 and 110.2 ka stepwise decreasing calcite contents from 110.2 ka to a minimum at 108.4 ka indicate another more pronounced cooling. This second minimum is relatively well constrained by the occurrence of tephra layer OT0702-9, which correlates with the marine X-6 tephra layer (Vogel et al. 2009). At Lago Grande di Monticchio Brauer et al. (2007) correlated a distinct decrease in arboreal pollen percentage centred around the occurrence of the X-6 tephra layer to the North Atlantic C24 cold event. Raising calcite contents from 108.4 ka until 103.3 ka to values similar to those during the LI indicate that climatic conditions during the interstadial of MIS 5.3 and the LI (MIS 5.5) at Lake Ohrid were similar. This is also suggested for Lago Grande di Monticchio (Brauer et al. 2007). From ca. 103.3 ka temperatures gradually decreased again at Lake Ohrid, such as documented in decreasing calcite values (Fig. 4). The period between 97.6 and 81.7 ka is due to the hiatus not documented in core Co1202.

## MIS 4, 3, and 2 (81.7–15 ka)

The last glacial climatic and environmental conditions between 81.7 and 15 ka, including MIS 4, 3, and 2, are represented by Lithofacies 1a and 1b sediments, which are dominated by clastic fine to medium sized silts, frequent occurrences of IRD, low amounts of OM, and negligible calcite contents (Figs. 3, 5). Marginal

variation in sedimentary composition indicates a relatively stable low productivity environment.

Overall, climatic and environmental conditions between 81.7 and 15 ka seem to have been comparable to MIS 6 at Lake Ohrid (Figs. 3, 4, 5). The occurrence of IRD throughout this succession indicates that the lake was at least partly ice covered during winter. Low amounts of OM and authigenic calcite (Fig. 5) indicate low productivity, likely related to low spring-summer surface temperatures, reduced supply and replenishment of nutrients and $Ca^{2+}$ and $HCO_3^-$ ions, and fostered decomposition of OM and dissolution of calcite by a well oxygenated water column and surface sediment. This general picture of rather cold climate conditions at Lake Ohrid corresponds relatively well to other paleoclimate reconstructions in the Mediterranean (Allen et al. 1999; Bar-Matthews et al. 1999; Martrat et al. 2004).

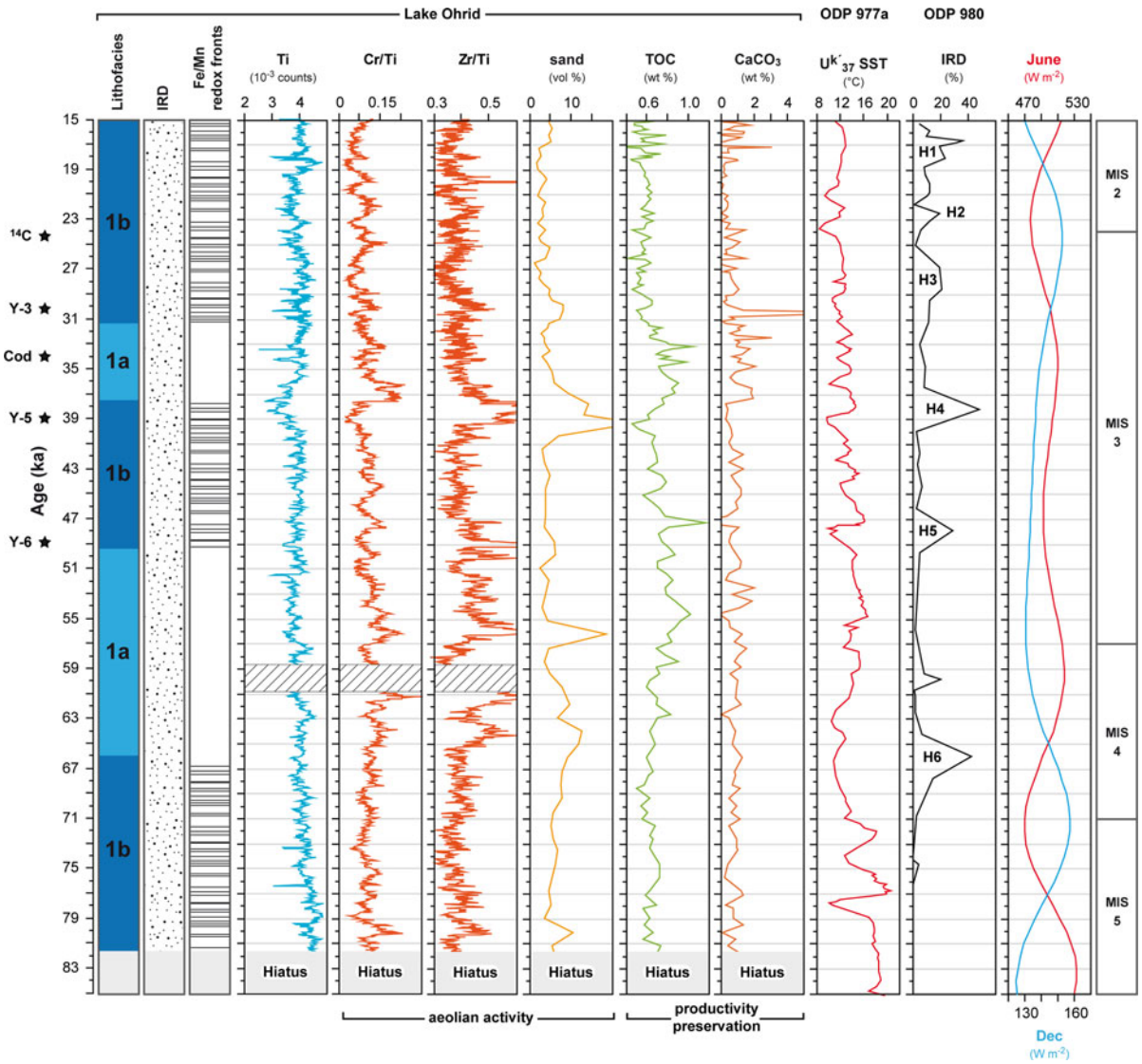


**Fig. 5** Lithofacies (legend shown in Fig. 3), IRD occurrences, Fe/Mn redox front occurrences, Ti intensities, Cr/Ti, Zr/Ti ratios, sand, TOC, $CaCO_3$ contents, and tephrochronological age control points (*asterisks*) of core Co1202 between ca. 81.7 and 15 ka. Also shown are $U_{37}^{K/}$ derived sea surface temperatures from ODP site 977a (Martrat et al. 2004), IRD contents from ODP site 980 (McManus et al. 1999), the local June and December (Dec) insolation (Berger and Loutre, 1991), and marine isotope stages (MIS; Bassinot et al. 1994). H6–H1 = Heinrich events 6–1

Low amplitude climatic and environmental variation can be inferred from faint changes in lithology and OM of core Co1202 (Fig. 5). The occurrence of Fe/Mn redox fronts in combination with minima of OM in Lithofacies 1b between 81.6–66.1, 46.7–38.2, and 31.7–15 ka point to enhanced mixis of the water column and significantly reduced productivity. The preservation of such redox fronts requires abrupt and recurrent changes of the sedimentation regime (Deike et al. 1997), and is probably related to abrupt and recurrent melt water pulses from local snowfields and ice caps in the Lake Ohrid catchment (Belmecheri et al. 2009). Other explanations include variations in vegetation cover and moisture supply, which are known to have been highly variable during this period in the Mediterranean (Allen et al. 1999). The interspersed Lithofacies 1a sediments, in contrast, are characterised by moderately increased amounts of OM and the absence of Fe/Mn redox fronts (Fig. 5), which suggests slightly enhanced productivity due to warmer temperatures and a more stable catchment during these periods. The record from Lake Ohrid only matches in part the minimum temperature inferred for the glacial maxima of MIS 4 (71–57 ka) and 2 (24–11 ka) and slightly warmer conditions during the MIS 3 interstadial (57–24 ka) (Bassinot et al. 1994; Fig. 5). Instead, phases of high OM content and absence of paleo-redox fronts correlate more closely with the local summer insolation pattern (Fig. 5; Berger and Loutre 1991).

Despite these rather moderate long-term climatic changes, short-term climatic variability and its imprint on the environment at Lake Ohrid is indicated by quasi-cyclic changes in the amount of fine sand and Cr/Ti and Zr/Ti ratios. Correlated maxima are centred on 80, 64, 62, 56, 50, 38, and 30 ka (Fig. 4), with the timing of the latter two periods being relatively well defined by tephra layers OT0702-6/Y-5 and OT0702-4/Y-3 (Vogel et al. 2009). The maxima during MIS 4, 3, and 2 point to stronger wind activity and/or enhanced erosion of sparsely vegetated soils. For periods around 38 and 30 ka a cold and dry climate with relatively sparse vegetation coverage in the Lake Ohrid catchment is confirmed by a multiproxy study on core Lz1120 from the southeastern part of the lake (Wagner et al. 2009). The cold and dry climate was tentatively correlated with Heinrich events 4 and 3. The sparse vegetation coverage around Lake Ohrid during these periods is probably partly related to enhanced erosion of soils in the catchment and led to the increases in fine sand-sized material and Cr/Ti and Zr/Ti ratios in core Co1202. Moreover, the timing of maxima in fine sand-sized material in core Co1202 at 64 and 50 ka corresponds relatively well with Heinrich events 6 and 5 at ca. 66 and 48 ka, respectively (McManus et al. 1999; Sánchez Goñi et al. 2000; Fig. 5). Maxima in fine sand, Cr/Ti and Zr/Ti occurring at 80, 62, and 54 ka (Fig. 5) indicate that cold and dry climate conditions persisted also during periods with less severe changes in Northern Hemisphere atmospheric circulation than those correlated to Heinrich events. More subtle quasi-cyclic fluctuations in the highly-resolved Cr/Ti and Zr/Ti ratios between 81.7 and 15 ka imply that climatic and environmental conditions were subject to recurrent changes throughout the last glacial at Lake Ohrid. A pattern of cyclic and recurrent changes in dust flux is also recorded in ice core records from Greenland and has been assigned to perturbations of the atmospheric circulation (Mayewski et al. 1997). Direct linkage of the pattern observed at Lake Ohrid to the pattern found in Greenland ice cores is, however, hampered by chronological uncertainties in both records.

## Termination I and the Holocene (15 ka–present)

Climatic and environmental conditions for the period from 15 ka to the present at Lake Ohrid are indicated by the occurrence of sediments assigned to Lithofacies 1c (362–264 cm) and 1b (264–246 cm) and 2a (246–222, 106–52, 25–0 cm) and b (222–106, 52–25 cm; Fig. 6). Based on distinct differences in sediment characteristics varying from clastic dominated sediments of Lithofacies 1b to calcite dominated sediments of Lithofacies 2b and varying strongly in grain-size distribution and OM amounts it can be inferred that climatic and concomitant environmental changes altered significantly during this period.

The period from 14.7 to 11 ka is characterised by abundance of IRD indicating cold winters with at least partial ice cover of the lake. A modest increase of OM and calcite content from 14.7 ka implies a stronger productivity due to higher spring-summer temperatures, which are probably related to the onset of warmer climate conditions in the Mediterranean during the Bölling/Alleröd interstadial complex (Allen et al. 1999; Martrat et al. 2004; Bordon

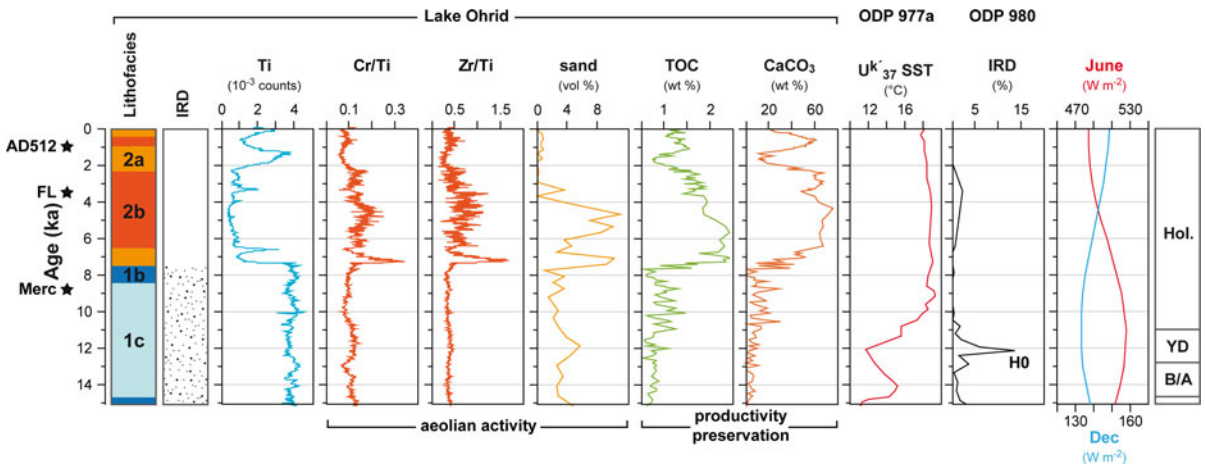


**Fig. 6** Lithofacies (Legend shown in Fig. 3), IRD occurrences, Ti intensities, Cr/Ti, Zr/Ti ratios, sand, TOC, $CaCO_3$ contents, and tephrochronological age control points (*asterisks*) of core Co1202 between ca. 15 ka and the present. Also shown are $U^{K'}_{37}$ derived sea surface temperatures from ODP site 977a (Martrat et al. 2004), IRD contents from ODP site 980 (McManus et al. 1999), the local June and December (Dec) insolation (Berger and Loutre 1991), and marine isotope stages (MIS; Bassinot et al. 1994). H0 = Heinrich event 0, B/A = Bölling/Alleröd, YD = Younger Dryas, Hol. = Holocene

et al. 2009). The subsequent Younger Dryas cooling between 12.7 and 11.6 ka (Allen et al. 1999; Martrat et al. 2004; Bordon et al. 2009) cannot be inferred with confidence from OM and calcite profiles at Lake Ohrid. However, the fine-sand and Cr/Ti, Zr/Ti maxima in core Co1202 centred at ca. 11.8 ka (Fig. 6) might be an expression of stronger wind intensities and enhanced erosion during the Younger Dryas, when the soils were dry and sparsely vegetated (Bordon et al. 2009).

From ca. 11 ka stepwise increases in OM and calcite content (Fig. 6) probably mark the beginning of warm Holocene climate conditions at Lake Ohrid, and is consistent with other paleoclimate reconstructions (Allen et al. 1999; Martrat et al. 2004; Lawson et al. 2004; Kotthoff et al. 2008; Bordon et al. 2009). It is not until 7.5 ka, however, that IRD disappears from the sediments of Lake Ohrid and that OM and in particular calcite contents rise to levels comparable to the LI. The presence of IRD is an expression of cold winters at Lake Ohrid during this period, which is in contrary to most other paleoclimate reconstructions in the region (Lawson et al. 2004; Bordon et al. 2009) except for the Lago Grande di Monticchio record where slightly colder winter temperatures are also indicated (Allen et al. 2002). Hence, a possible explanation might be a stronger influence of cold polar air outbreaks during winter between ca. 11–7.5 ka. The Lake Ohrid valley might have particularly been affected by these polar air masses due to its relatively northward location and its N–S direction. Relatively low calcite contents could be related to complete mixis of the water body and dissolution. In addition, depletion of $Ca^{2+}$ and $HCO_3^-$ ion concentration in the water column might be related to a stronger contribution of surface runoff and direct precipitation and to a wetter early Holocene at Lake Ohrid as also suggested by other paleoclimate studies in the Mediterranean (Aritztegui et al. 2000; Zanchetta et al. 2007; Roberts et al. 2008).

The subsequent rapid increase in OM and calcite contents and absence of IRD from 7.5 ka (Fig. 6) indicate that spring-summer temperatures likely increased and that the lake remained ice-free during winter. A contemporaneous increase in fine sand content and Cr/Ti and Zr/Ti ratios between 7.3 ka and 6.9 ka indicates an increase in wind activity. Calcite contents reach maximum concentrations comparable to the LI at ca. 6.4 ka. Since reconstructed temperatures (Martrat et al. 2004; Allen et al. 1999; Bordon et al. 2009) do not exhibit major fluctuations from ca. 7.5 ka to the present it can be assumed that highest contents of calcite in Lake Ohrid between 6.4 and 2.4 ka are due to a generally warm but also drier climate as also suggested by other paleoclimate reconstructions in the region (Roberts et al. 2008; Kotthoff et al. 2008). A decrease in OM and calcite concentrations between ca. 4.3 and 3.0 ka, coincident with a similar decrease in core Lz1120 from the

southeastern part of the lake (Wagner et al. 2008), might indicate reduced primary productivity due to lower temperatures, in line with temperature reconstructions from the surrounding Adriatic (Sangiorgi et al. 2003) and Aegean Seas (Rohling et al. 2002).

Between 3.0 and 2.4 ka calcite content rises again to levels comparable to the preceding period, while OM contents remain relatively low. Low OM during this period could be due to enhanced mixis rather than reduced productivity, which would have also strongly affected calcite precipitation. A rapid decrease in both OM and calcite content commencing at 2.4 ka is probably related to anthropogenic deforestation in the catchment of Lake Ohrid (Wagner et al. 2009), which possibly led to increased erosion and surface runoff, enhanced mixis and decomposition (Wagner et al. 2009), and/or lower productivity.

OM and calcite contents increase again from 1,000 to 800 years, falling thereafter until ca. 100 years (Fig. 6). This peak might result from enhanced productivity due to increased nutrient supply from Lake Prespa via karst aquifers (Wagner et al. 2009), when the lake level of Prespa was 6 m lower than today (Matzinger et al. 2006b). This also suggests a more arid climate around this period. Hence, the relatively high amounts of calcite between ca. 1,000 and 300 years in core Co1202 could additionally be a result of stronger evaporation and lower contribution of surface runoff and direct precipitation to the hydrological budget of Lake Ohrid. Rapid environmental changes for the period from ca. 1,000 years to the present might be the expression of climate and concomitant environmental changes of the "Medieval Warm Period" and the following "Little Ice Age" (Wagner et al. 2009). However, it cannot be excluded that the results of anthropogenic activity during the late Holocene (Wagner et al. 2009) have overridden any signature of natural climate variability.

## Conclusions

Core Co1202 from the northeastern part of Lake Ohrid provides substantial information on climatic variability and their imprint on the local hydrology, limnology, and catchment dynamics back to MIS 6. Sedimentary characteristics at Lake Ohrid differ significantly between glacial and interglacial stages due to distinct and complex changes in the hydrological budget, catchment dynamics, and limnology.

The environment of penultimate (MIS 6, 136–127.3 ka) and last glacial (MIS 4, 3, 2; 81–11 ka) periods at Lake Ohrid seem to have been relatively similar and was characterised by a cold climate and a relatively stable low productivity. Subtle, long-term climatic and environmental changes are indicated with coldest conditions during the periods between 81.6–66.1, 46.7–38.2, and 31.7–15 ka and slightly warmer interspersed periods at Lake Ohrid. Quasi-cyclic fluctuations in fine-sand contents and Cr/Ti and Zr/Ti ratios are a result of particularly cold and dry climate conditions, which led to increased erosion of sparsely vegetated soils and an increase in wind activity around 129, 64, 50, 38, 30, and 11.8 ka. These periods were tentatively correlated to Heinrich events 11, 6, 5, 4, 3, and 0, respectively.

Climatic and environmental conditions during MIS 5.5, 5.4, and 5.3 (127.3–96 ka) and the Holocene (11 ka-present) at Lake Ohrid are characterised by elevated OM and calcite contents indicating enhanced productivity. Relatively low winter temperatures during the early Holocene (11–7.5 ka) are indicated by the occurrence of IRD. Relatively humid climate conditions likely prevailed between ca. 11–6.5 ka. Several abrupt events during the interglacials seem to be related to short-term cooling events. During the late Holocene climatic and environmental variability is probably blurred by significant anthropogenic impact on the catchment at Lake Ohrid.

Overall, Lake Ohrid's sediments sensitively record evidence for long and short-term climatic and environmental changes. Variations in sediment composition are generally synchronous at different locations within the lake basin. Therefore, Lake Ohrid provides an excellent archive of climatic and environmental variability, which can be well correlated to other records in the Mediterranean and the northern Hemisphere, but owing to the spatial variability of this climatically complex region also exhibit some local peculiarities.

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