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Mollusca stable isotope record of a core from Lake Frassino, northern Italy: hydrological and climatic changes during the last 14 ka

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Abstract: A core retrieved from Lake Frassino (northern Italy) provided evidence of palaeohydrological change in this area during the last 14 ka. Lithological, malacological and, in particular, stable isotope composition of freshwater shells allowed the delineation of the main phases of lake evolution between the Lateglacial and Holocene. Lateglacial conditions were drier than the Holocene, although a wetter period was inferred before *c.* 14 ka. According to the oxygen isotope composition of freshwater shells, the Holocene showed a clear bipartition. The first part, which lasted from ~9100 to 7000 yr BP, was drier and was followed by a rapid increase in humidity at ~7000–6800 yr BP. Between ~6800 and 5000 yr BP, there were wetter and more stable conditions, as indicated by smaller $\delta^{18}O$ oscillations of *Pisidium* and *Valvata* shells. From *c.* 5000 to 2600 yr BP the record is characterized by larger fluctuations, which may indicate that short-term and particularly pronounced alternation of wet and dry periods occurred.

Key words: Stable isotopes, lake level, hydrological change, climatic change, Lateglacial, Holocene, Northern Italy, molluscs.

Introduction

Northern Italy is an interesting area on which to focus palaeoclimate studies concerning the Lateglacial to Holocene periods since it is in a key position between central Europe, dominated by the effect of North Atlantic circulation, and central and southern Italy where Mediterranean climatic conditions dominate (eg, Zolitschka *et al.*, 2000; Bolle, 2003). The perialpine area of northern Italy is studded with numerous small and shallow lakes that originated during the last glacial retreat, whose sediments can provide natural archives for investigating the palaeoenvironmental evolution of this area. Small lakes are extremely sensitive to changes in amount of rainfall, source of moisture and changes in local temperature (Eicher and Siegenthaler, 1976; Fritz *et al.*, 2000; Yu *et al.*, 2002). These changes can be investigated through the stable isotope composition of biogenic (eg, freshwater shells) and authigenic precipitates, providing important proxy data for palaeohydrological and palaeoclimatic reconstruction on land (Stuiver, 1970, Eicher and Siegenthaler, 1976; Siegenthaler and Eicher, 1986; von Grafenstein *et al.*, 1994, 1999; Leng and Marshall, 2004).

Isotopic records from the Italian lakes are relatively sparse (Robinson et al., 1993; Bonadonna and Leone, 1995; Chondrogianni *et al.*, 1996, 2004; Zanchetta *et al.*, 1999; Longinelli *et al.*, 2000; Baroni *et al.*, 2001; Belis and Aritzegui, 2004) and mostly concentrated on central Italy. These records have yet to provide a clear picture of the hydrological changes that occurred during the Lateglacial and the Holocene in the Italian peninsula. Moreover, it is difficult to compare the isotopic profiles with the $\delta^{18}O$ record obtained from lakes located in central and northern Europe (eg, Eicher *et al.*, 1981; von Grafenstein *et al.*, 1994; Ahlberg *et al.*, 1996, 2001; Whittington *et al.*, 1996) and the absence of common patterns may indicate regional and subregional climatic and hydrological differences.

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This paper illustrates the results of isotopic investigations of a core retrieved near the shore of 'Laghetto del Frassino' (hereafter Lake Frassino, Figure 1) and the inferred hydrological changes that occurred during the Lateglacial and Holocene. This study can also provide information on past lake level changes in the nearby Lake Garda (Figure 1C; Baroni, 1985), which may have overflowed into Lake Frassino, and to constrain the timing of the last glacial retreat in the area.

Site description

Lake Frassino is located south of Lake Garda (45°26′08″N, 10°39′48″E) at 74 m a.s.l. (Figure 1). It is a small shallow lake (surface 0.30 km^2; maximum depth: *c*. 15 m; volume *c*. 2.4 10^6 m^3; catchment area 4.14 km^2; Barbato, 1987) directly fed by local rainfall and small inlet streams (Figure 1B). Lake water temperatures are available for 1981–82. According to Barbato (1987), thermal stratification is apparent during summer and late autumn with a thermocline located at *c*. 10–12 m, which disappears during winter. The maximum surface temperature was recorded in August (*c*. 23°C), whereas the lowest was in December–January (*c*. 4°C). The lake occupies an inner position with respect to the last glacial terminal moraine (Cremaschi, 1987; Figure 1C), and it presumably developed soon after the last glacial retreat. A temperate subcontinental climate is present in the area today (Pinna, 1970). The mean annual temperature of the area is 12°C, and the mean annual precipitation is 869 mm (years 1959–1982, data from the weather station at Peschiera, Figure 1C). The highest mean temperature, 21.7°C, occurs in July, and the minimum, 2.4°C, in January (Figure 2). The highest precipitation occurs in October and November (mean 90 mm) and the lowest during January and February (mean 56 and 53 mm, respectively, Figure 2). The mean yearly isotopic composition of local meteoric precipitation can be estimated as *c*. $-8.0‰$, and *c*. $-52‰$ for δ^{18}O and δD, respectively (Longinelli and Selmo, 2003).

Materials and methods

Three cores were retrieved using a continuous coring system (diameter of core barrel 100 mm) on the lake shore during 1994 (S1, S2 and S3, Figure 1B). Core S3, the deepest, supplied 15 sections of 1.5 m each to a depth of 25.5 m where the bottom of the lake sediment was reached, intercepting gravelly deposits at *c*. 23 m. Therefore, S3 was selected for detailed investigation. The core was sectioned longitudinally, described and sampled soon after coring operations. The 15 core sections were further cut into 50-cm sections and stored at 4°C at the Dipartimento di Scienze della Terra di Pisa. Along the core, 179 samples were collected for potential stable isotope investigation and malacological analyses. Each lithological change was sampled just below and above the boundary. Samples were usually 1 cm thick (*c*. 80 cm^3). The sediment was disaggregated in 10% H_2O_2 solution and washed through 500 and 250 μm sieves, rinsed several times in deionized water and then dried. Shells were manually hand-picked under a binocular microscope, identified and counted. The most common shells along the profile were selected for stable isotope analyses. They were the bivalve *Pisidium* spp. and the gastropod *Valvata piscinalis*. *Pisidium* spp. are typical infaunal species (eg, Girod *et al*., 1980) whereas *Valvata piscinalis* is epifaunal, preferring to live on submerged macrophytes (eg, Girod *et al*., 1980) in well-oxygenated water. *Pisidium* shells are a mixture of aragonite and calcite (von Grafenstein *et al*., 1999), while *Valvata piscinalis* shells are pure

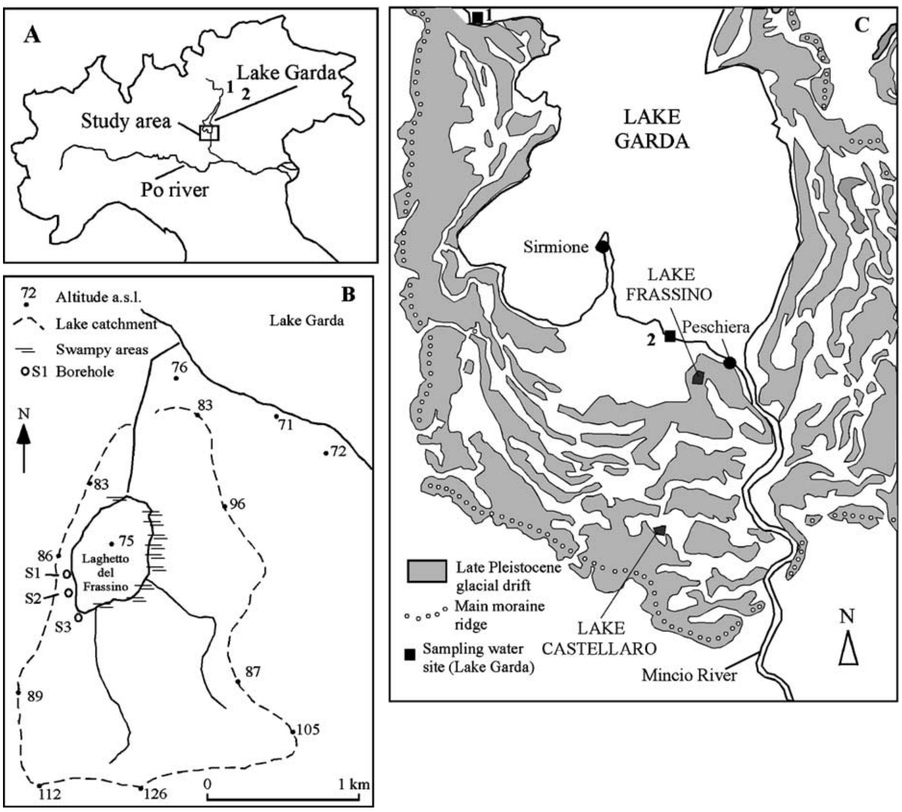

Figure 1 (A) Location map of the study area. The location of other northern Italian isotopic records discussed in the text are also noted: 1, Lake Terlago; 2, Grotta di Ernesto Cave. (B) Sketch of Lake Frassino and the position of the cores. (C) Simplified geological map of the study area (after Cremaschi, 1987). Sites of water collection at Lake Garda are also shown: 1, Salo; 2, Tiglio (Sirmione)

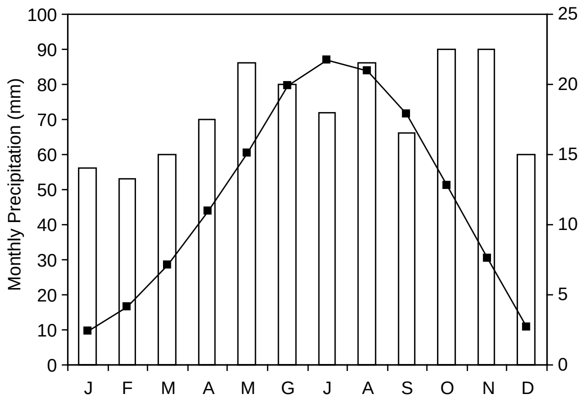

Figure 2 Climatic data from the weather station of Peschiera (mean of the years 1959–1982).

Table 1 Stable isotope ($\delta^{18}O$ and $\delta^{13}C$) data from freshwater shells collected at the bottom of Lake Frassino (*c.* 1.5–2 m depth)

Species	$\delta^{13}C_{PDB}$ ‰	$\delta^{18}O_{PDB}$ ‰
Valvata piscinalis	– 7.43	– 3.84
Valvata piscinalis	– 6.67	– 2.09
Valvata piscinalis	– 4.27	– 3.34
Valvata piscinalis	– 4.69	– 3.64
Bithynia tentaculata	– 6.29	– 4.68
Bithynia tentaculata	– 8.10	– 3.12
Bithynia tentaculata	– 8.28	– 4.31
Physa fontinalis	– 10.51	– 4.26
Physa fontinalis	– 11.33	– 2.45
Lymnaea sp.	– 8.18	– 4.27
Lymnaea sp.	– 9.05	– 4.61

aragonite (Linz and Müller, 1981). The X-ray diffraction (XRD) of a bulk sample, obtained from individuals of *V. piscinalis* from different depths, yielded pure aragonite. The best-preserved shells were selected and cleaned in an ultrasonic bath, dried and powdered, then ashed for approximately 8 h in a low-temperature oxygen plasma to remove organic contaminants. Isotopic analysis was performed with a PRISM II mass spectrometer connected to an ISOCARB automated CO_2-extraction device with a H_3PO_4 bath at 90°C. The isotopic results are reported in the well-known δ‰-notation relative to V-PDB. The standard deviation (1σ) on replicate samples was ±0.05‰ and ±0.06‰ for carbon and oxygen, respectively. Most the samples comprised several individuals of the same species: 2 to 5 for *Valvata piscinalis* and 3 to 20 for *Pisidium* spp. To assess the isotopic variability ($\Delta\delta = \delta_{max} - \delta_{min}$) within samples, single shells of *V. piscinalis* were also separated and analysed. The mean variability was ±1.3‰ and ±1.1‰ for $\delta^{13}C$ and $\delta^{18}O$, respectively.

The isotopic composition (δD and $\delta^{18}O$) of water of three lakes (Frassino, Castellaro and two localities at Lake Garda; Figure 1C) were monitored during the years 2002–2003. Lake waters were collected using polyethylene bottles, carefully sealed after sampling to avoid evaporation and stored at 4°C before analysis. The oxygen isotope composition was measured by means of the water-CO_2 equilibration technique (Epstein and Mayeda, 1953). The δD measurements were carried out by reducing the water to hydrogen by passage over hot zinc at about 520°C, according to the procedure described by Coleman *et al.* (1982). The standard deviation (1σ) of $\delta^{18}O$ measurement was ±0.05‰ and ±1.0 ‰ for δD measurements. The isotopic results are reported in δ‰ units versus V-SMOW. All isotopic results are available from the authors on request.

In addition, at Frassino, shells of *Valvata piscinalis*, *Bithynia tentaculata*, *Physa fontinalis* and *Lymnaea* sp. were sampled at the lake bottom and isotopically analysed (Table 1). These shells still contained remnants of organic tissue, and were therefore presumed to have died recently.

Core chronology was established through seven AMS ^{14}C-dates (Table 2, calibrations were performed using the program Calib Rev 5.0.1; Reimer *et al.*, 2004), five of which were obtained from the organic fraction of bulk samples, one from seeds and one from *Pisidium* spp. shells. All the dates are in stratigraphic order except the lowest sample consisting of *Pisidium* shells, which has a very large standard error because of the low amount of carbon. The chronology of the core was, therefore, established assuming constant sedimentation rate between consecutive dates and rejecting the *Pisidium* values (Figure 3). The age of the bottom of the isotopic record was calculated assuming the same sedimentation rate between the last two ages. Calculation yielded an age of *c.* 16 ka BP. Because the bulk samples for ^{14}C-dating may comprise mixed lacustrine and terrestrial organic matter, and lacustrine organic matter can be affected by unknown hard-water and reservoir effects, the core chronology should be considered with care. Throughout the paper, uncalibrated ^{14}C dates are used (^{14}C BP) for the Frassino record. Other records discussed report calibrated ages as cal. BP.

Results

Lithology

The core is characterized by five main lithological changes (Figure 4).The basal section (2335–2550 cm) comprises gravels and silty sands, which probably represent glacial deposits. From 2335 to 930 cm the core mainly comprises siliclastic muds with discontinuous millimetric and submillimetric laminations and rare intercalation of fine sands. Upward of this interval there is a visible increase in organic matter, which, at *c.* 884 cm, passes abruptly to a few centimetres-thick peat layer. High organic matter content continues up to 840 cm. From 840 cm to 172 cm marls prevail, with local increases in

Table 2 ^{14}C-dating obtained from the core S3. Calibrations were performed using the program Calib Rev 5.0.1 (Reimer *et al.*, 2004). Radiocarbon dates were supplied by Geochron Laboratory Krueger Enterprise Inc. Cambridge, Massachusetts (GX-); Isotrace Radiocarbon Laboratory, Toronto (TO-); Centrum voor Isotopen Onderzoek, Groningen (GrA-)

Sample no.	Depth (cm)	Material	^{14}C age (yr BP)	^{14}C age cal. BP	$\delta^{13}C_{VPDB}$‰
GX-25108	305–306	Peat	3650±40	3990–3900	– 29.4
GrA-20424	775–776	Bulk organic matter	7350±70	8200–8040	–
GX-25107	864–865	Peat	9090±50	10 190–10 270	– 24.6
GrA-20423	884–885	Bulk organic matter	10 220±70	12 070–11 800	–
GrA-20422	955–956	Bulk organic matter	11 000±70	12 980–12 870	–
GrA-20342	1024–1026	Seeds	13 990±70	16 880–16 470	–
TO-7720	1073–1075	*Pisidium* spp.	12 300±2330	–	–

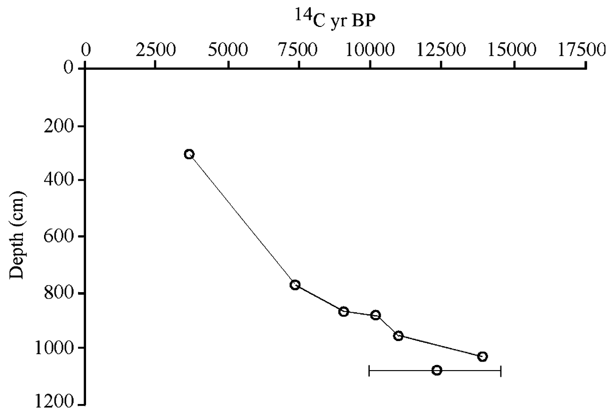

Figure 3 ^{14}C-dates versus depth. Where the standard error is not shown it means that it is smaller than the size of the symbols

organic matter. The very top of the core is made up of an alternation of organic marls and peat.

Non-marine molluscs

The shell record is discontinuous with some sterile sections. The most important gaps occur between 0 and 170 cm (peat and organic mud deposits), 235 and 248 cm, 266 and 271 cm, 873 and 887 cm, 1041 and 1064 cm and from 1124 to 2334 cm. The number of specimens (Figure 5) fluctuates from 0 to a maximum of 136 per sample. On average a higher number of shells was collected from *c.* 680–860 cm and from 170 to 200 cm. The association of *Pisidium* spp. and *Valvata piscinalis* dominates throughout the record, but the interval from 870 to 1050 cm is almost entirely dominated by *Pisidium* spp. However, *Valvata piscinalis* and *Pisidium* spp. sometimes have an antipathetic behaviour, which may indicate lake-level oscillations or changes in bottom conditions. The almost constant occurrence of *Pisidium* spp. and/or *Valvata piscinalis* indicate a permanent lake at least some metres deep for most of the core (Girod *et al.*, 1980). For instance *V. piscinalis* usually lives in water depths of 2–5 m (Kerney, 1999) The *Pisidium* spp. and *Valvata piscinalis* assemblages are replaced between *c.* 200 and 270 cm by few specimens of *Lymnaea peregra*, *Armiger crista*, *Segmentina nitida* and *Gyraulus albus.* A few specimens of *Bithyinia tentaculata*, *Lymnaea peregra*, *Valvata cristata* and *Segmentina nitida* occur mainly discontinuously upward of 700 cm. *Armiger crista* and *Gyraulus albus* also occur below 700 cm, but *A. crista* is more frequent upward. *Succinea oblonga* occurs discontinuously throughout the core; however, no particular trends are appreciable. *S. oblonga* is a typical hygrophilous species, which favours lake shores (Kerney and Cameron, 1987), and its rare presence is indicative of transportation from nearby shores. The low numbers of species at most levels, and the low number of individuals (except for *Pisdium* and *Valvata piscinalis*) in the sequence mean that the usual diagram of molluscan percentages by depth has not been drawn. Instead, total molluscan numbers for each level are shown in Figure 5.

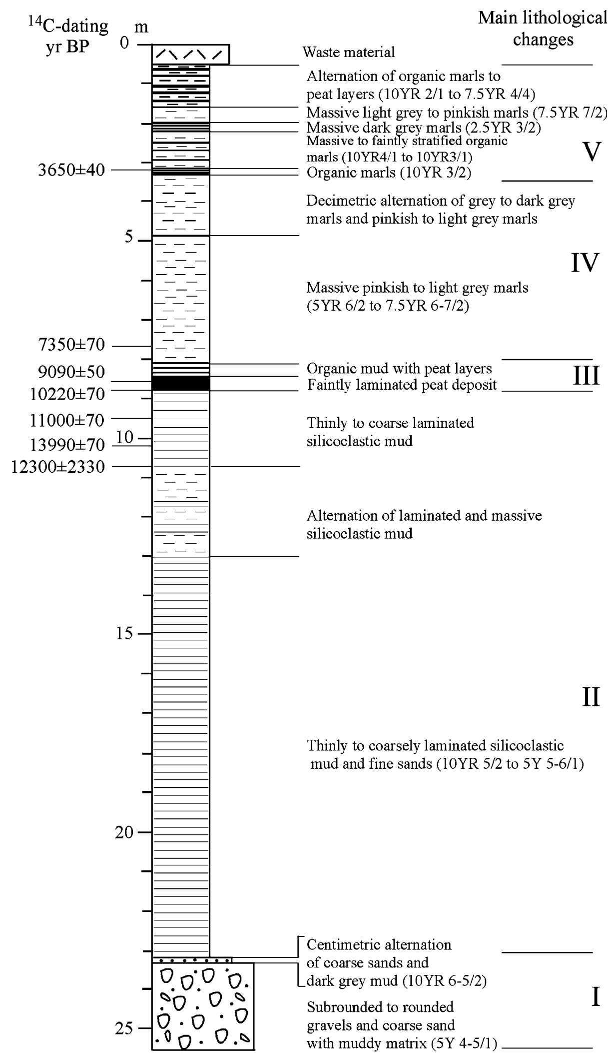

Figure 4 Stratigraphy and lithology of core S3. The main lithological changes (I–V) discussed in the text are also shown

Stable isotopes

The mean isotopic composition of Frassino Lake water is $-3.6 \pm 0.6‰$ and $-32.6 \pm 2.2‰$ for oxygen and hydrogen, respectively. In the $\delta^{18}O$–δD diagram (Figure 6) the isotopic composition of Frassino and Castellaro lakes fits with a regression line:

$$\delta D = 4.5\delta^{18}O - 16.5 \ (R^2 = 0.97) \quad (1)$$

Equation (1) represents an evaporation line with respect to the local meteoric water line (LMWL) with the equation (Longinelli and Selmo, 2003):

$$\delta D = 7.7\delta^{18}O + 9.4 \ (R^2 = 0.98) \quad (2)$$

The intersection point between equations (1) and (2) has values $-8.1‰$, and $-53‰$ for $\delta^{18}O$ and δD, respectively. This is in perfect agreement with the estimation of the mean isotopic composition of local rainfall (Longinelli and Selmo, 2003) and confirms that the local meteoric precipitation is the main source of water for both lakes. The differences found in the two lakes are due to varying seasonal degrees of evaporation, a common phenomenon in a chain of lakes with different sizes, shapes and volumes (eg, Turner *et al.*, 1983; Gonfiantini, 1986; Yu *et al.*, 2002). In contrast Lake Garda has less variable and lower $\delta^{18}O$ and δD (Figure 5). The greater volume of Lake Garda and the lower isotopic composition of the main inlet river (Sarca River) that has its catchment in the Alps explain these differences. The Sarca River has $\delta^{18}O$ values ranging from *c.* $-10‰$ to $-15‰$, with the most ^{18}O-depleted values measured in spring because of snow melt (A. Longinelli, unpublished data, 2000). However, longer residence time of Lake Garda waters produces detectable evaporation of surface water. Surprisingly, the $\delta^{18}O$ measured for Lake Garda surface water is almost perfectly homogeneous ($-7.4 \pm 0.2‰$; A. Longinelli, unpublished data, 2000) for most of the nearshore samples, with the only exception being the north-

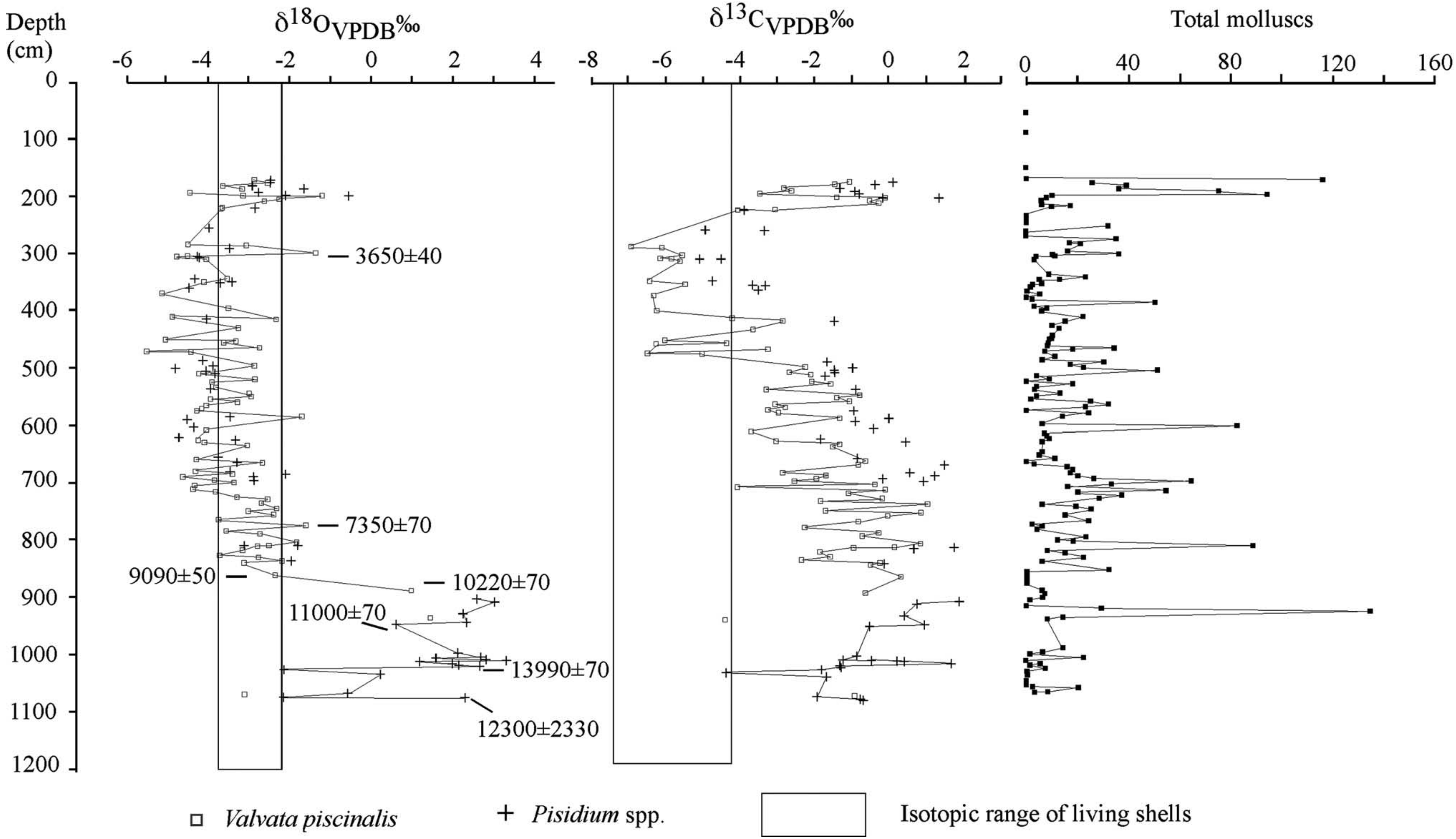

Figure 5 $\delta^{18}O$, $\delta^{13}C$ records and total number of freshwater shells of core S3. The isotopic range measured on living shells of *Valvata piscinalis* is also shown for comparison

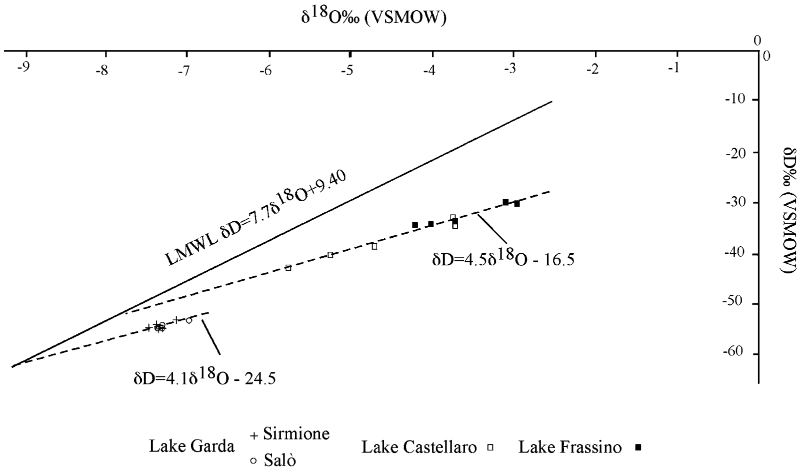

Figure 6 δ^{18}O versus δD diagram of water samples from Lake Garda, Lake Castellaro and Lake Frassino. The local meteoric water line is from Longinelli and Selmo (2003)

ernmost section. In δ^{18}O–δD space these values fit with a regression line with the equation:

$$\delta D = 4.1\delta^{18}O - 24.5\ (R^2 = 0.78) \quad (3)$$

The intercept of this straight line and the LMWL correspond to a δD of *c.* -63.1‰ and δ^{18}O of *c.* -9.4‰. These values may be close to the initial isotopic composition of the surficial Lake Garda water, which originated by mixing from river water (essentially Sarca water) and meteoric precipitation from the catchment area (several streams with variable flows).

Given the variability of the isotopic composition of the Lake Frassino water, the δ^{18}O values found for the shells of the same species collected at the bottom of the lake can be interpreted as those of individuals that died recently but likely lived over different years with different degrees of evaporation of lake water and/or at slightly different temperature (eg, Jones *et al.*, 2002).

Along the core record the mean δ^{18}O difference between coexisting *Valvata* and *Pisidium* is only -0.4 ± 1.0‰. This small difference could be due to a different vital offset, water δ^{18}O and/or differences in the temperature of shell calcification. Since *Valvata* is, on average, slightly depleted in ^{18}O with respect to *Pisidium*, this may rule out mineralogy as one of the factors in that aragonite is enriched in ^{18}O relative to calcite by *c.* 0.6‰ at 25°C (Tarutani *et al.*, 1969). Ambient temperature may depend either on different habitat depth, if the lake has a significant temperature gradient or a different time of shell calcification (Chaix *et al.*, 1982; Lemeille *et al.*, 1983; von Grafenstein *et al.*, 1994, 1999, 2000; White *et al.*, 1999; Jones *et al.*, 2002). However, it is apparent that the δ^{18}O record of Lake Frassino is little biased by species-dependent or temperature effects resulting from different lake depth of calcification (Figure 5). Moreover, the variability of isotope ratios at a given depth is of the order of that found in living species. As discussed extensively by Jones *et al.* (2002), the differences found in each stratigraphic level and also in the recent shells may depend on: (i) variability of isotopic composition of water and temperature in different years; (ii) sedimentation rate; and (iii) possible reworking. The number of shells used to form samples may also be important (Jones *et al.*, 2002). Despite the large fluctuations in the isotope record and the variability found in two samples and in living individuals, the δ^{18}O record seems to retain a genuine environmental signal, and both *Valvata* and *Pisidium* can be, with small margin for error, used interchangeably to infer the general δ^{18}O trend.

Calcification temperatures of the living *Valvata piscinalis* shells, calculated using the equation (Hoefs, 2004):

$$T(°C) = 17 - 4.52(\delta^{18}Os - \delta^{18}Ow) + 0.03(\delta^{18}Os - \delta^{18}Ow)^2 \quad (4)$$

are around 12–13°C, which, in case of negligible vital offset, are consistent with shell growth occurring during most of the year, even if higher growth occurred during warmer months (Chaix *et al.*, 1982; von Grafenstein *et al.*, 1999). In equation (4) the δ^{18}Os is the mean value of living shells and δ^{18}Ow is the mean value of lake water. Since the *Valvata piscinalis* shells are of pure aragonite, the δ^{18}Os values were corrected by 0.6‰ (Tarutani *et al.*, 1969).

As far as the shell δ^{18}O record is concerned, at least two main intervals are readily recognized (Figure 5): a lower interval comprising 890 to 1077 cm ($\sim$10 300–16 200 BP) with the higher δ^{18}O values (mean 1.4 ± 1.6‰) and an upper one spanning 173 to 863 cm ($\sim$2600–9100 BP) with a mean of -3.4 ± 1.1‰. The transition occurs between 863 and 905 cm ($\sim$9100–10 300 BP) where a δ^{18}O shift of *c.* 4‰ occurs. However, within both intervals there are some sections characterized by distinct δ^{18}O values. In particular, in the lower interval there is an important decrease of δ^{18}O of several per mill at 1029–1075 cm ($\sim$14 100–16 100 BP). This section with lower δ^{18}O values shows a rapid termination with an ^{18}O-enrichment of $\sim$2‰ in less than three centuries.

The main feature of the δ^{18}O record of the upper interval is the evident lowering of the δ^{18}O values between 863 and 730 cm (mean values: -2.8 ± 0.6‰) with respect to the interval above 730 cm (mean values: -3.6 ± 0.9‰). The transition between the two sections occurred in about two centuries between $\sim$6800 and 7000 yr BP. In general the upper interval (above 863 cm) is consistent with the δ^{18}O values found in the recent *Valvata piscinalis* specimens (Figure 5), with greater similarity between 863 and 730 cm.

The δ^{13}C of freshwater shells depends mainly on the carbon isotope composition of the dissolved inorganic carbon (DIC) into the lake, when vital effects are negligible (Fritz and Poplawski, 1974; Burchardt and Fritz, 1980; von Grafenstein *et al.*, 1999). However, the δ^{13}C values of the shells of different species are controlled by the microenvironment in which the species live and also by the presence of vital offsets (eg, von Grafenstein *et al.*, 1999) as also shown by the δ^{13}C of modern shells collected at Lake Frassino (Table 1). Effectively, the mean δ^{13}C difference between *Valvata piscinalis* and *Pisidium* along the core is -1.6 ± 1‰. This difference cannot be due to different mineralogy, in that aragonite-calcite fractionation is

1.7±0.4‰ and independent of temperature from 10 to 40°C (Romanek *et al.*, 1992). However, a coherent $\delta^{13}C$ trend for both species is apparent, at least above 863 cm. If the *Valvata piscinalis* record is considered, which is the most continuous above 863 cm, four different sections are present (Figure 5). These are between 863 and 701 cm (~9100–6800 BP; mean $\delta^{13}C$: −0.9±1.2‰), 696 and 497 cm (~6700–5200 BP; mean $\delta^{13}C$: −2.2±0.9‰), 471 and 285 cm (~5000–3500 BP; mean $\delta^{13}C$: −5.5±1.2‰), and 221 and 173 cm (~3000–2600 BP; mean $\delta^{13}C$: −1.9±1.4‰). The transition between 863–701 cm and 696–497 cm intervals matches fairly well the most evident changes in the $\delta^{18}O$ record, whereas the interval between 471 and 285 cm corresponds with a phase in which the $\delta^{18}O$ record shows very large oscillations. In the $\delta^{18}O$–$\delta^{13}C$ diagram the shells below 863 cm (that is, older than *c.* 10000 yr BP) mainly plot on a very distinct field with respect to those above 863 cm (Figure 7).

Discussion

The oldest age of the core record (13 990±70 BP GrA-20342) agrees with other evidence indicating that the withdrawal of the southern alps glaciers began before the onset of the Lateglacial Insterstadial (Cavallin *et al.*, 1997; Baroni *et al.*, 2001). In fact, the basal gravels (Figure 4), possibly representing an ice-contact deposit linked to the stationing of a glacial tongue during the first phase of the retreat after the last glacial maximum, are older than the estimated age of ~16 000 BP for the base of the isotopic record. Up to *c.* 16 000 BP, lake sediments were sterile, indicating that lake waters were unable to sustain mollusc populations. Whether this was due to very localized conditions, for example high water turbidity (Gilberson, 1980; Ložek, 1986), or conditions under which winter ice persisted into the summer and prevented the atmospheric oxygenation of the water, thus making molluscan life impossible (Ökland, 1990), is difficult to establish. It is possible that this part of the core records deposition in a glacial-dammed lake. Mollusc assemblages progressively developed after 16 000 BP, indicating the onset of favourable conditions, which may point to progressive climatic improvements in the aftermath of the Lastglacial period. The first part of the mollusc record is dominated, up to 863 cm, by *Pisidium* spp., which probably can be considered one of the first mollusc colonizers of the Alpine and Peri-alpine lakes formed after the withdrawal of the glaciers (Baroni *et al.*, 2001). Similar to the mollusc record, the most profound change in the core lithology occurred, marked by the passage from laminated siliciclastic muds to marl (Figure 4), at about the transition between the Lateglacial and the Holocene (*c.* 10 000 BP). This lithological change is separated by a transitional interval characterized by a visible increase in organic matter and a decrease in the number of molluscs with the development of a peat layer, which indicates a phase of lowering of lake level (eg, Magny, 2001). This is followed by a lake transgression, which marked the onset of the marl sedimentation. At this time, the number of molluscs increased significantly. On average, the mollusc numbers are highest between *c.* 860 and 675 cm. The age of peat deposition may indicate a short dry phase characterized by a drop in lake level at the beginning of the Holocene. It is interesting to note that at the very beginning of the Holocene in the Swiss Plateau, the Jura Mountains and the northern French Pre-Alps a short dry phase has been recognized, both on stable isotope profiles (von Grafenstein *et al.*, 2000) and lake level records (Magny, 2001).

The $\delta^{18}O$ values of living shells (Table 1) and those of the waters (Figure 6) provide evidence that today the lake isotopic composition is mainly controlled by the rate of evaporation and/or the supply of meteoric precipitation. It is reasonable to assume that the $\delta^{18}O$ record should be mainly interpreted as changes in the hydrological budget of the lake. The degree of ^{18}O-enrichment of lake water resulting from evaporation depends on air relative humidity, temperature differences between surface water and air, rate of vapour removal (ie, wind fetch and speed) and the rate of water recharge (ie, meteoric precipitation) (Craig and Gordon, 1965; Gonfiantini, 1986). Higher $\delta^{18}O$ values of lake water, and therefore higher $\delta^{18}O$ values of shells, should correspond to phases of reduced precipitation and/or increase of evaporation rate (ie, drier period), while lower $\delta^{18}O$ values should correspond to an increase in the amount of precipitation and/or reduction of the evaporation of lake water (ie, wetter period). This can be complicated by temperature changes and by the changes in the isotopic composition of meteoric precipitation, which can act in concert to enhance or partially suppress the isotopic effects linked to changes in the rate of evaporation (Dansgaard, 1964; Rozanski *et al.*, 1993; Fricke and O'Neil, 1999). For instance, if drier periods are characterized by lower lake temperatures this

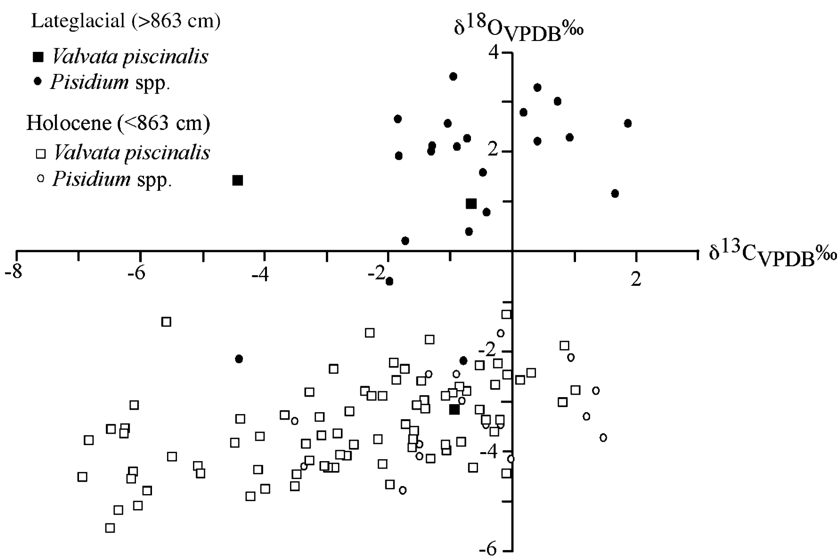

Figure 7 $\delta^{18}O$ versus $\delta^{13}C$ diagram for the core record. Note that shells older than 10 000 yr BP (*c.* below 863 cm) occupy a different field respective to those younger than 10 000 yr BP

produces a further ^{18}O-enrichment in shells, and the opposite if humid periods are characterized by higher lake temperatures. Moreover, substantial seasonal changes in the amount of rainfall feeding the lake can be an additional complicating factor. Today, summer precipitation has higher $\delta^{18}O$ values compared with winter (Borsato *et al*., 2003).

Despite these complications the mollusc $\delta^{18}O$ seems to record first a wetter phase from ~16 000 to 14 100 BP, followed between ~13 900 and 10 400 BP by drier climate conditions (Figure 5). However, the resolution of this interval precludes a detailed reconstruction of Lateglacial hydrological changes. In contrast, the Holocene is characterized by wetter conditions. However, the first part, lasting from ~9100 to 7000 BP, was drier and was followed by a rapid establishment of wetter conditions at ~7000–6800 BP. In this period the $\delta^{18}O$ values are highly consistent with the values of recent *Valvata piscinalis* shells (Figure 5). Between ~6800 and 5000 BP, there were probably wetter and more stable conditions, as suggested by the smaller $\delta^{18}O$ oscillations of *Pisidium* and *Valvata*.

From *c*. 5000 to 2600 BP the record is characterized by larger fluctuations, which may indicate that short-term and particularly pronounced alternation of wet and dry periods occurred. The most pronounced $\delta^{18}O$ peaks indicating dry conditions are centred at *c*. 5800, 4500, 3600 and 2800 BP. Aridification phases with comparable chronologies have been identified by pollen analyses in Italy (Ramrath *et al*., 2000; Sadori and Narcisi, 2001), southeastern Spain and France (Jalut *et al*., 2000) and by lake level changes (Magny *et al*., 2002). However, a detailed correlation is precluded by the generally low chronological resolution of these events at both Lake Frassino and in the pollen records.

A strong decrease in the $\delta^{18}O$ values at the Lateglacial to Holocene transition has been observed in other Mediterranean records, such as Konya Basin and Eski Acigöl (Turkey) (Leng *et al*., 1999; Roberts *et al*., 2001) and Lake Zeribar (Iran) (Stevens *et al*., 2001). For these records changes in the water balance of the lake (Eski Acigöl and Konya Basin), and changes in seasonality of precipitation (Lake Zeribar) have been invoked to explain the observed shift. Independently of the origin of the $\delta^{18}O$ shifts, this is a peculiar characteristic of the Mediterranean area compared with central Europe, where many records show an increase of $\delta^{18}O$ at the beginning of the Holocene, mainly interpreted as the effect of increased temperatures (eg, Eicher and Siegenthaler, 1976; von Grafenstein *et al*., 1994, 2000).

The apparent strong bipartition of the $\delta^{18}O$ record during the Holocene is noteworthy. The palaeohydrological records from studies of lake level oscillations from the Mediterranean region in general suggest a clear bipartition during the Holocene period (eg, Magny *et al*., 2002 and references therein). The first half is generally characterized by relatively wet conditions in comparison with the second part, marked by progressively drier conditions. This general observation is regionally and chronologically articulated (Magny *et al*., 2002) and in partial conflict with the Lake Frassino $\delta^{18}O$ record. The higher $\delta^{18}O$ values between *c*. 9000 and 7000 BP suggest drier conditions during this time, and possibly a lower lake level, consistent with the higher number of molluscs found in this interval (Figure 5). Indeed, higher numbers of shells are usually found in shallower waters (eg, Magny, 2001). However, lower $\delta^{18}O$ values after *c*. 7000 BP suggest wetter conditions, possibly with a trend toward drier conditions at *c*. 3500 BP. On average this is consistent with the variation in the total number of freshwater molluscs. Drier conditions during the first part of the Holocene have been inferred from the $\delta^{18}O$ record of a stalagmite from Grotta di Ernesto (Trentino, NW Italy, Figure 1A); the transition toward wetter conditions occurs at *c*. 8000 cal. BP (McDermott *et al*., 1999). It is notable that the $\delta^{18}O$ record preserved at Lake Steisslinger (southwest Germany; Mayer and Schwark, 1999) shows overall drier conditions during the first part of the Holocene, with a very prominent decrease in the $\delta^{18}O$ values at *c*. 8000 cal. BP (Mayer and Schwark, 1999). The $\delta^{18}O$ record of Lake Terlago (Figure 1A) also shows higher $\delta^{18}O$ in the first part of the Holocene (Baroni *et al*., 2001). However, this record, although potentially important for local correlation, has a very low chronological resolution.

Evident bipartition of the Holocene has also been found in the $\delta^{18}O$ record of Eski Acigöl (Roberts *et al*., 2001) and Lake Zeribar (Stevens *et al*., 2001). In both records, lower $\delta^{18}O$ values occur in the early to middle Holocene compared with the Late Holocene, which is not as evident in the Frassino $\delta^{18}O$ record.

The $\delta^{18}O$ record allows us to reject any possibility that during the last 16 ka BP Lake Garda overflowed the catchment of Lake Frassino. If this event(s) occurred in the past it would have to have been accompanied by a dramatic decrease (now at least 4‰) in the isotopic composition of lake water, which does not seem consistent with the available data. Any Garda overflow of significance into Lake Frassino must have occurred prior to the estimated age of 16 000 BP.

As far as the isotopic composition of carbon is concerned, interpretation is complicated by the fact that in lacustrine systems the $\delta^{13}C$ of the DIC is controlled by several processes (eg, Leng and Marshall, 2004) including: exchange with atmospheric CO_2, respiratory activity of aquatic plants, mineralization of organic matter and DIC isotopic composition of the inflowing water, which depends on the dissolution of old carbonate rocks and leaching of soil CO_2. A particularly pronounced ^{13}C-enriched DIC can originate by the dissolution of old carbonates and/or equilibration with atmospheric CO_2 (eg, Leng *et al*., 1999), enhanced by CO_2 outgassing linked to strong evaporation (Talbot, 1990). A DIC generated by the oxidation of organic matter produces generally ^{13}C-depleted values (Deines, 1980). When the amount and $\delta^{13}C$ values of external DIC supply do not change significantly through time, primary productivity and processes of recycling (eg, oxidation versus methanogenesis) of organic matter become important in governing the evolution of the lake DIC (McKenzie, 1985; Hollander and McKenzie, 1991; Curry *et al*., 1997; Hollander and Smith, 2001).

Although the shell $\delta^{13}C$ record shows brief and relatively large fluctuations, the most important changes appear crudely associated with changes in the $\delta^{18}O$ record (Figure 7), which may indicate an indirect climatic control of the evolution of the lake DIC. During the Lateglacial conditions persist in which high evaporation dominates and the $\delta^{13}C$ of *Pisidium* spp. is higher during the interval of higher $\delta^{18}O$. This is consistent with increased residence time of lake water (ie, increase in the evaporation/inflow ratio) during the drier period, in which there could be an increase in the extent to which lake-water DIC is isotopically equilibrated with atmospheric CO_2 (Hammarlund *et al*., 2003). This may also indicate inefficient recycling of organic matter into the lake, and possibly low primary productivity, during drier periods coupled with low input of soil-leached CO_2 from the surrounding catchment. Notably, some $\delta^{13}C$ lake records in Europe show higher $\delta^{13}C$ values during Lateglacial, eg, Lake Steisslinger (southwest Germany, Mayer and Schwark, 1999), Lough Gur (western Ireland, Ahlberg *et al*., 2001) and Lundin Tower (eastern Scotland, Whittington *et al*., 1996). Relatively high $\delta^{13}C$ values at Frassino persist during the first part of the Holocene up to *c*. 7000 yr BP. In agreement with $\delta^{18}O$ values, this could be linked with a still high residence time of lake water during the first part of the Holocene, which enhanced isotopic equilibration with atmospheric CO_2. We also suggest that the increase of biological productivity in the lake in the first part of the Holocene, which preferentially removes ^{12}C from the CO_2 pool, may have maintained the $\delta^{13}C$ of the DIC at a higher level. However, a direct comparison with Lateglacial conditions cannot be made. As suggested by Figure 7, two very different sets of lake conditions seem to be present between the Lateglacial and the Holocene. After *c*. 7000 BP the $\delta^{13}C$ record indicates a lake DIC progressively ^{13}C-depleted, reaching a minimum between *c*. 5000 and 3500 BP. This indicates a progressive relative increase of oxidized organic matter into the system, and between 5000 and 3500 BP the lake seems to possess a DIC similar to today, as shown by the $\delta^{13}C$ values of the living *Valvata piscinalis* (Figure 5).

During the Holocene, Lough Gur (Ahlberg *et al*., 2001) and Lake Steisslinger (Mayer and Schwark, 1999) show a comparable pattern of progressive decrease of $\delta^{13}C$ values, with the lower values around *c*. 5000 to 3000 BP. A similar pattern could be inferred for Lundin Tower (Whittington *et al*., 1996), although there the Holocene chronology is not supported by ^{14}C-dating. Between 3000 and 2600 BP the DIC in the lake experienced a new ^{13}C-enrichment, which is consistent with an increase of residence time during a drier phase, as indicated by the $\delta^{18}O$ record.

During the Holocene a positive correlation between $\delta^{18}O$ and $\delta^{13}C$ is apparent (Figure 7), although with a low degree of correlation ($R^2 = 0.27$). As noted by Talbot (1990), shells usually show a lower degree of covariance compared with authigenic carbonates (ie, marls), probably because bulk marl samples better average the lake condition than shells. Drummond *et al*. (1995) suggested that in temperate regions, for lakes with size and depth comparable with those of Lake Frassino, the isotopic covariance of lake carbonates is directly influenced by the change in the length of summer months. In temperate regions summer meteoric precipitation has higher $\delta^{18}O$ (eg, Fricke and O'Neil, 1999) than winter, and summer organic productivity enriches epilimnic DIC in ^{13}C. Thus, climate changes toward longer summers could result in greater proportions of warm-month meteoric precipitation, longer duration of warm-month productivity and long-term enrichment in carbonate ^{18}O and ^{13}C. This model is not necessarily in contrast with the increase of evaporation (higher in warm months), increase in residence time and increase of the extent of isotopic equilibration with atmospheric CO_2 during drier phases. However, these effects could be complicated by changes in the seasonal contrast, in terms of the amount of meteoric precipitation between summer and winter months. Today this contrast is not very strong (Figure 2) in the study area, but it could have been different in the past.

Conclusions

The Lake Frassino record highlights the existence of strongly different lake conditions between Lateglacial and Holocene. These differences are recorded by changes in lithology, mollusc assemblage and, most importantly, clear $\delta^{18}O$ changes. The $\delta^{18}O$ changes were mainly interpreted as changes between drier and wetter periods, even if multiple factors acted in concert to influence the isotopic signal. During the Lateglacial, drier conditions seem to prevail, with a wetter phase recorded between *c*. 16 000 and 14 100 BP.

The first part of the Holocene (*c*. 9000–7000 BP) was drier than the period between *c*. 7000 and 2500 BP. The wettest conditions were present between *c*. 7000 and 5000 BP. The upper part of the record indicates more unstable climate conditions with significant oscillation between wetter and drier phases. However, many questions on the climate evolution of the area remain. Of paramount importance for future research should be the recovery of a detailed record of the Lateglacial and an investigation of centennial-scale oscillation during the Holocene, a crucial point in current Quaternary research (Mayewski *et al*., 2004). There are still too few studies on proxy climate records from Northern Italy.

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