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ORIGINAL PAPER

Arctic freshwater ostracods from modern periglacial environments in the Lena River Delta (Siberian Arctic, Russia): geochemical applications for palaeoenvironmental reconstructions

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Abstract The aim of this study is to describe ostracods from freshwater habitats in the Siberian Arctic in order to estimate the present-day relationships between the environmental setting and the geochemical properties of ostracod calcite. A special focus is on the element ratios (Mg/Ca, Sr/Ca), and the stable isotope composition (δ^{18}O, δ^{13}C), in both ambient waters and ostracod calcite. The most common species are *Fabaeformiscandona pedata* and *F. harmsworthi* with the highest frequency in all studied waters. Average partition coefficients D(Sr) of *F. pedata* are 0.33 ± 0.06 (1σ) in females, and 0.32 ± 0.06 (1σ) in males. A near 1:1 relationship of δ^{18}O was found, with a mean shift of $\Delta_{\text{mean}} = 2.2‰ \pm 0.5$ (1σ) to heavier values in ostracod calcite of *F. pedata* as compared to ambient waters. The shift is not dependent on $\delta^{18}\text{O}_{\text{water}}$, and is caused by metabolic (vital) and temperature effects. Temperature-dependence is reflected in the variations of this shift. For ostracod calcite of *F. pedata* a vital effect as compared to inorganic calcite in equilibrium was quantified with 1.4‰. Results of this study are valuable for the palaeoenvironmental interpretation of geochemical data of fossil ostracods from permafrost deposits.

Keywords Freshwater ostracods · Element ratios · Stable isotopes · Lena River Delta · Siberian Arctic

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Introduction

Freshwater ostracods are crustaceans, usually less than 3 mm long, with a bivalved carapace made of low magnesium-calcite. During their ontogeny, the individuals run through nine moult stages (Kesling 1951). After each moult, ostracods calcify new shells within a short time, probably within a few days (Chivas et al. 1983). The ions for the calcite formation are incorporated directly from the ambient water at this time (Turpen and Angell 1971). Therefore, element ratios of magnesium, strontium and calcium (Mg/Ca, Sr/Ca) as well as stable isotope ratios of oxygen (δ^{18}O) and carbon (δ^{13}C) in ostracod calcite are related to the geochemistry of the ambient water (e.g., Chivas et al. 1986; Xia et al. 1997a, b; von Grafenstein et al. 1999). The understanding of these relationships is a prerequisite for interpreting geochemical information in fossil ostracod calcite for palaeoenvironmental reconstructions.

Due to the sensitivity of freshwater ostracods to environmental changes and the high durability of their remains in lacustrine sediments, ostracods serve as good indicators for palaeoclimatic reconstructions (e.g., Anadón et al. 2006; Poberezhnaya et al. 2006; Xia et al. 1997c). The ecology of freshwater ostracods is defined by water chemistry, water temperature and additional habitat parameters such as area, water depth, and water permanency and type. Environmental changes influence the diversity of freshwater ostracods as well as the morphology and the geochemical composition of the ostracod shells (e.g., Griffiths and Holmes 2000). Coupled element and stable isotope measurements in fossil ostracod calcite are being increasingly used in palaeoenvironmental reconstructions of temperature, salinity and productivity (e.g., De Deckker and Forester 1988; Griffiths and Holmes 2000). Whereas fossil and modern ostracod fauna and their ecology in mid-latitudes are relatively well known due to numerous investigations, there are only a few records concerning freshwater ostracods in Siberia, or in (sub-) Arctic permafrost regions (e.g., Alm 1914; Bunbury and Gajewski 2005; Pietrzeniuk 1977; Semenova 2005).

Numerous multidisciplinary studies have focused on permafrost deposits with well preserved remains of palaeoindicative fossils and have highlighted their potential and significance as Quaternary palaeoclimate archives in the Siberian Arctic (e.g., Schirrmeister et al. 2002a, b, 2003; Hubberten et al. 2004), especially since other long-term records such as lake sediment cores or inland glacier ice cores are rare or not available in this vast region. Various palaeoproxies in the frozen deposits, such as pollen (Andreev et al. 2002), rhizopods (Bobrov et al. 2004), plant macro-fossils (Kienast et al. 2005), insects (Kuzmina and Sher 2006) and mammal bones (Sher et al. 2005), as well as stable isotope records of ground ice (Meyer et al. 2002), have already been used, while ostracods were introduced as a valuable palaeoindicator only recently by Wetterich et al. (2005).

Still, knowledge about the ecology and biology of Arctic freshwater ostracods needs improvement to apply modern analogues to fossil records. Thus, the key question is how do ostracod associations and/or the geochemistry of their valves reflect the natural setting of a modern aquatic environment in the periglacial Arctic, which is characterised by widely distributed polygons and thermokarst lakes in a tundra landscape. It should be mentioned that this study presents the species distribution and life conditions of ostracods at the sampling time. Nevertheless, our results increase the value of using freshwater ostracods from permafrost deposits as palaeoindicators and, when combined with planned geochemical studies on element ratios (Mg/Ca, Sr/Ca) and stable isotopes (δ^{18}O, δ^{13}C), will enable us to interpret signals from fossil ostracod calcite in this region.

Study area and types of water bodies

The Lena River has the largest delta in the Arctic, located at the Laptev Sea shore between Taimyr Peninsula and the New Siberian Islands (Fig. 1a). The delta covers an area of about 32,000 km^2, where more than 1,500 islands of various size were formed by a network of rivers and channels (e.g., Are and Reimnitz 2000). The islands are composed of Quaternary sediments. They are subdivided into three terraces of different ages and height levels above the modern flood plain of the Lena River (e.g., Grigoriev 1993; Schwamborn et al. 2002).

The modern climatic conditions of the Lena River Delta are extreme. Short and cool summers follow long and very cold winters. The mean July air temperature varies between +4°C and +8°C and the mean January air temperature varies between −36°C and −32°C (Atlas Arktiki 1985). The mean annual air temperature averages −14°C (Kunitsky 1989). Only 4 months of the year are mean air temperatures above 0°C. Consequently, shallow lakes and ponds are covered with ice and frozen to the bottom during at least eight to nine months of the year, which circumscribes the period of active growth for annual aquatic organisms such as ostracods in this region. The mean annual precipitation (about 200 mm) is low, but still higher than evaporation, because of the predominance of freezing temperatures through the year (Kunitsky 1989). Data from 2002 by an automatic soil and meteorology measurement station on Samoylov Island (Friedrich and Boike 1999; Wille et al. 2003) recorded mean air temperatures 0.5 m above the ground and mean soil temperature 0.15 m below the ground as the following: in May (T_{Air} = −9.0°C; T_{Soil} = −10.5°C), in June (T_{Air} = 6.2°C; T_{Soil} = −0.1°C), in July (T_{Air} = 10.7°C;

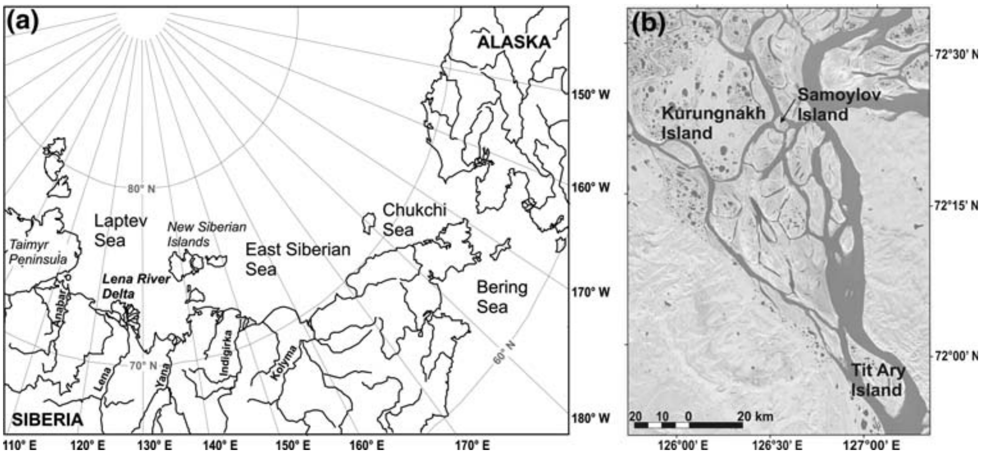


Fig. 1 (**a**) Location of the Lena River Delta on the Laptev Sea coast in northeast Siberia; (**b**) the study area in the southern part of the delta (Satellite image provided by Statens Kartverk, UNEP/GRID-Arendal and Landsat 2000)

T_{Soil} = 6.1°C) and in August (T_{Air} = 10.0°C; T_{Soil} = 7.4°C). The precipitation totalled 65 mm (J. Boike, AWI Potsdam, unpublished data).

The entire Lena River Delta is located in the zone of continuous permafrost, with a permafrost thickness of 400–700 m in the western part and 300–500 m in the eastern part (Kondrat'eva and Solov'ev 1989). The occurrence of permafrost is expressed on the landscape microrelief as widespread patterned ground dominantly formed by ice wedge polygons in different stages of development. The formation of lakes, ponds, and other water bodies on islands of the Lena River Delta is directly connected to permafrost processes like ice wedge growth, thermokarst, and thermoerosion, as well as to the fluvial dynamics of the Lena River on the flood plain.

The patterned ground of the polygonal tundra is dominated by ponds and thaw lakes (Fig. 2). In winter, initial frost cracks are generated by rapid temperature drops below 0°C (Fig. 2a). In spring, these cracks are filled by melt water, which freezes immediately because of the negative temperatures in the ground. Ice wedges, formed by successive annual cycles of this process, grow in a polygonal pattern. The polygon rim is usually higher than the polygon centre. Intrapolygon ponds (Fig. 2b) are situated in these so-called low-centre polygons (French 1996). Later, the degradation of the polygon rims and changes in the hydrological regime may cause the formation of so-called high-centre polygons (French 1996), which are often accompanied by interpolygon ponds and thaw lakes (Fig. 2c–d). Shallow intrapolygon ponds have a water depth between about 0.5–1 m, but are characterised by different diameters (Meyer 2003). The size of intrapolygon ponds can be up to 30 m in diameter, depending on the polygon in which they occur. Interpolygon ponds delineate the location of polygon rims and are underlain by ice wedges. Thaw lakes are also shallow with a water depth of up to 1.5 m and cover areas of up to several hundred m^2. Thermokarst processes cause extensive melting of the underlying permafrost and large depressions thus develop, which often form lakes. Thermokarst lakes occur over areas up to several km^2 with water depths up to 5 meters. On the lower floodplain, shallow cut-off river branches (up to 1 m water depth) expand during spring flooding and form stream-oriented shallow depressions.

Recent ostracods were found during limnological investigations in 40 shallow lakes and ponds on Samoylov (72° 22′ N, 126° 28′ E), Kurungnakh (72° 20′ N, 126° 10′ E), and Tit Ary Islands (71° 58′ N, 127° 04′ E) in the southern part of the Lena River Delta (Fig. 1b). Here, we present data from 23 lakes and ponds situated on Samoylov (on the first Lena River terrace and lower flood plain) and Kurungnakh Islands (on the third Lena River terrace), where ostracods were found in sufficient numbers for further geochemical analyses. The studied waterbodies included three intrapolygon ponds, three interpolygon ponds, 13 thaw lakes and one river branch on Samoylov Island as well as one intrapolygon pond

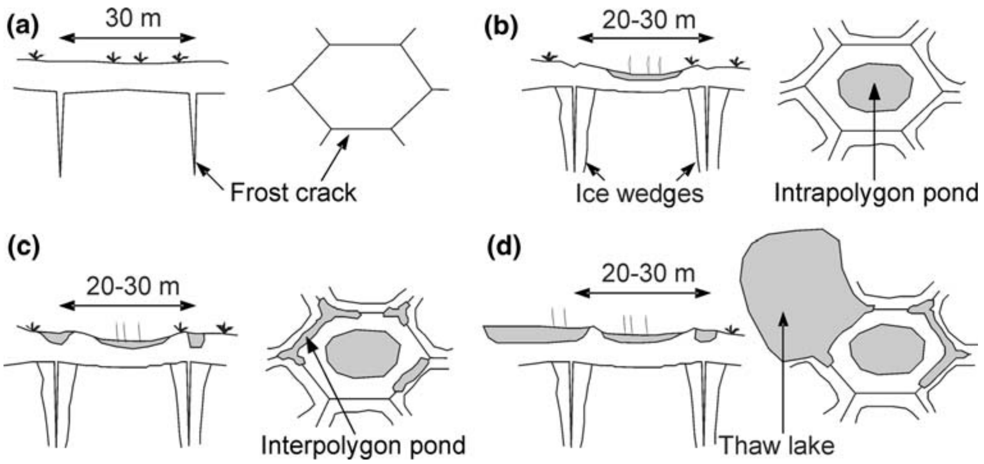


Fig. 2 The formation of ice wedges and the resulting changes on the landscape surface with different types of shallow waterbodies depending on the polygon degradation state in the Lena River Delta (modified after Meyer 2003). (**a**) Juvenile polygon type with very small height differences between polygon wall and the centre; no waterbody; (**b**) Mature low-centre polygon type with height differences between 0.5 m and 1 m between polygon wall and the centre; intrapolygon ponds develop in poorly drained sites; (**c**) Polygon type of initial degradation with height differences between 0.5 m and 1 m between polygon wall and the centre; interpolygonal ponds on the polygon wall are present as triangular ponds in the triple junctions or elongated ponds along the frost crack; (**d**) Polygon type of final degradation with height differences of up to 1.5 m between polygon wall and polygon centre; thaw lake of variable size and polygonal structure are present

and two thermokarst lakes on Kurungnakh Island (Table 1).

Materials and methods

Hydrochemical variables such as pH, the content of nutrients (NH_4, NO_2, NO_3, PO_4) and oxygen (O_2) were analysed during the fieldwork by means of a compact laboratory (Aquamerck). All water samples and measurements were performed at the margin of the water bodies at water depths of about 0.5 m, where near-bottom water was taken or measured, respectively. Water temperature and electrical conductivity were measured with a conductivity meter (WTW Cond 330i). Water was sampled for ionic (Ca, K, Mg, Sr, Na, Cl, SO_4, HCO_3) and stable isotope (δD, δ^{18}O, δ^{13}C) analyses. Samples for cation analyses were acidified with HNO_3, whereas samples for anion analyses and residue samples were preserved by freezing until analysis. Before conservation, samples for cation and anion analyses were passed through a cellulose-acetate filter (0.45 µm pore size). Upon return to the main laboratory, the element (cation) content of the waters was analysed by Inductively Coupled Plasma-Optical Emission Spectrometry (ICP-OES, Perkin–Elmer Optima 3000 XL), while the anion content was determined by Ion Chromatography (IC, Dionex DX-320). The hydrogen carbonate concentrations (alkalinity) of the waters were determined by titration with 0.01 M HCl using an automatic titrator (Metrohm 794 Basic Titrino).

The lake water samples for oxygen and hydrogen stable isotope analysis (δD, δ^{18}O) were stored cool and later analysed by equilibration technique (Meyer et al. 2000) using a mass-spectrometer (Finnigan MAT Delta-S). The reproducibility derived from long-term standard measurements is established with 1σ better than ±0.1‰ (Meyer et al. 2000). All samples were run at least in duplicate. The values are expressed in delta per mil notation (δ, ‰) relative to the Vienna Standard Mean Ocean Water (VSMOW). The water samples for carbon isotope analysis (δ^{13}C) on total dissolved inorganic carbon (TDIC) were preserved by freezing until analysis using a mass-spectrometer (Finnigan MAT 252). The reproducibility derived from standard measurements over a 1-year period is ±0.1 ‰ (1σ). TDIC was extracted from lake water with 100% phosphoric acid in an automatic preparation line (Finnigan Gasbench I) coupled online with the mass-spectrometer. All samples were

Table 1 Location, type and hydrochemical characteristics of the studied lakes and ponds. The sample sites are arranged by water types as in Fig. 6. The water type identification follows the key: intrapolygon pond → intrapolygon; interpolygon pond → interpolygon; thaw lake → thaw lake; thermokarst lake shore → thermokarst and river branch of the Lena River → river branch

Sample Nº	Locality	Date yy-mm-dd	Water type	Sample depth [m]	Cond* [µS/cm]	pH	O_2 [mg/l]	T_{Air} [°C]	T_{water} [°C]	Mg ppm	Sr ppm	Ca ppm	HCO_3 ppm	Na ppm	Cl ppm	K ppm	SO_4 ppm
SAM-13	Kurungnakh	02–08–15	Intrapolygon	0.5	27.0	6.5	5.9	13.3	13.6	1.39	0.03	2.68	13.4	1.04	1.90	<0.3	<0.1
SAM-21	Samoylov	02–08–21	Intrapolygon	0.5	105.6	7.5	9.0	11.9	11.4	4.31	0.03	5.25	67.5	0.81	2.22	0.46	<0.1
SAM-23	Samoylov	02–08–21	Interpolygon	0.5	98.8	7.5	8.3	15.8	13.9	7.09	0.06	11.22	60.8	1.68	2.50	0.56	<0.1
SAM-25	Samoylov	02–08–25	Interpolygon	0.5	94.3	7.5	9.4	6.1	6.2	7.01	0.06	10.86	65.5	1.79	1.86	0.32	<0.1
SAM-30	Samoylov	02–08–27	Intrapolygon	0.5	106.6	7.5	10.6	10.5	8.0	7.23	0.06	8.87	66.4	1.37	2.21	0.76	<0.1
SAM-37	Samoylov	02–08–30	Interpolygon	0.5	93.3	7.2	11.3	7.9	9.3	6.88	0.06	11.07	61.4	1.68	2.61	0.43	<0.1
SAM-44	Samoylov	02–09–03	Intrapolygon	0.5	70.8	7.5	9.0	7.6	7.0	4.22	0.04	6.43	39.3	1.20	1.86	0.59	<0.1
SAM-01	Samoylov	02–08–02	Thaw lake	0.5	90.4	7.1	9.8	10.4	11.4	4.68	0.05	7.62	50.3	2.17	3.45	0.60	0.91
SAM-17	Samoylov	02–08–19	Thaw lake	0.5	64.9	7.5	8.4	11.1	12.0	3.80	0.04	8.07	38.3	0.88	1.20	0.56	<0.1
SAM-19	Samoylov	02–08–20	Thaw lake	0.5	53.1	7.5	8.8	13.5	12.5	2.84	0.04	6.31	31.8	1.53	2.35	0.47	0.73
SAM-24	Samoylov	02–08–21	Thaw lake	0.5	254.0	7.0	5.3	16.5	15.3	14.51	0.25	36.63	148	2.29	1.98	1.83	<0.1
SAM-26	Samoylov	02–08–25	Thaw lake	0.5	78.5	7.5	10.6	8.6	7.7	5.48	0.05	8.19	52.1	1.57	2.16	0.99	<0.1
SAM-27	Samoylov	02–08–25	Thaw lake	0.5	109.4	7.5	10.0	8.5	8.3	7.85	0.08	12.30	66.4	1.77	2.74	0.79	<0.1
SAM-28	Samoylov	02–08–26	Thaw lake	0.5	122.6	7.6	8.8	9.8	7.8	8.21	0.08	14.53	77.5	1.87	2.81	1.00	<0.1
SAM-29	Samoylov	02–08–26	Thaw lake	0.5	110.8	7.5	11.0	17.0	10.3	7.19	0.08	13.86	70.6	1.55	2.31	0.96	<0.1
SAM-32	Samoylov	02–08–27	Thaw lake	0.5	113.2	7.5	10.8	14.2	9.8	8.29	0.08	13.74	78.6	1.55	2.20	0.91	<0.1
SAM-33	Samoylov	02–08–29	Thaw lake	0.5	96.9	7.5	9.7	8.0	6.7	6.68	0.06	10.83	56.7	1.34	1.79	0.79	<0.1
SAM-34	Samoylov	02–08–29	Thaw lake	0.5	107.7	7.5	9.5	6.7	6.1	8.48	0.07	11.80	71.5	1.67	2.63	0.78	<0.1
SAM-40	Samoylov	02–09–01	Thaw lake	0.5	77.8	7.5	11.2	8.7	5.9	5.26	0.05	8.64	48.6	1.46	2.73	0.76	<0.1
SAM-41	Samoylov	02–09–01	Thaw lake	0.5	79.6	7.3	9.4	6.3	6.6	5.34	0.05	8.83	50.4	1.45	2.09	0.48	<0.1
SAM-12	Kurungnakh	02–08–15	Thermokarst	0.5	28.0	7.2	7.7	11.6	13.4	1.50	0.02	3.18	16.7	0.66	0.92	0.37	<0.1
SAM-38	Kurungnakh	02–08–31	Thermokarst	0.5	109.2	7.5	9.8	6.1	7.8	2.80	0.06	8.73	66.9	0.80	2.44	0.39	0.28
SAM-14	Samoylov	02–08–18	River branch	0.5	86.6	7.5	8.8	13.0	12.6	3.29	0.08	10.77	41.6	4.26	6.39	0.59	3.96

* Electrical conductivity

run at least in duplicate. The values are expressed in delta per mil notation (δ, ‰) relative to the Vienna Pee Dee Belemnite standard (VPDB).

Samples from the upper layer of bottom sediments (up to 5 cm) were analysed for nitrogen, organic and total carbon contents by CNS-Analyser (Elementar Vario EL III), as well as for grain-size distribution by Laser-Granulometry (Coulter LS 200).

Living ostracods were collected using an exhaustor (Viehberg 2002), then preserved in 70% alcohol and finally counted and identified under a binocular microscope (Zeiss SV 10) by soft body and valve characteristics described in Alm (1914), Bronshtein (1947), Neale (1969), and Meisch (2000). In samples with sufficient numbers of living ostracods, the most common species were prepared for element (Mg, Sr, Ca) and stable isotope (δ^{18}O, δ^{13}C) analyses. Additionally, subfossil valves from the upper layer of the bottom sediments were analysed in order to relate these data to that of the ostracods caught alive. The valves of these ostracods were cleaned by removing the soft body under the binocular microscope, and then washed in distilled water and air-dried. The subfossil valves were picked from the wet sieved 250 μm fraction of the upper layer of bottom sediments under the binocular microscope. Only clean valves of adult specimens were used for analysis. Particles adhered to valves were removed with a fine brush. Prior to analysis, up to four single valves from the same sample were weighed using an electronic micro-balance (Sartorius micro) in order to check the reliability of the $CaCO_3$ Nominal Shell Weight (NSW). The $CaCO_3$ NSW is calculated from total Ca concentrations of each sample solution (Chivas et al 1986). The single valve samples were placed in a reaction vial and dissolved in 30 μl of 20% HNO_3 (Baker Ultrex). Afterwards, 3 ml of distilled water were added. For analysis of Ca, Mg, and Sr contents we used Inductively Coupled Plasma-Optical Emission Spectrometry (ICP-OES, Varian Vista-MPX) at the Research Centre for Geosciences Potsdam (GFZ Potsdam, Germany). The ICP-OES was calibrated with three multi-element standards prepared by mono-element standard solutions for ICP (Alfa Aesar Specpure 1,000 μg/ml). Standard solution 1 contained 1 ppm Ca, 0.02 ppm Mg, and 0.01 ppm Sr. The concentrations in standard solutions 2 and 3 were twice and three times higher, respectively. Three determinations were made from each sample to check machine precision. Contaminant (blank) concentrations in the solvent acid were analysed for each batch of 10 samples to determine detection limits of the measurements. The detection limits in solution (3σ above background in μg/l or ppb, e.g., Doerfel 1966) are 1.51 for Ca (wavelength 422.673 nm), 0.05 for Mg (279.553 nm), and 0.03 for Sr (407.771 nm). Standard 1 provided a consecutive reference for each batch of 10 samples analysed, and confirmed an internal 1σ error of less than ±2.3% for Ca, ±1.0% for Mg, and ±1.2% for Sr. The results for Mg and Sr are expressed as μg/g (ppm) in calcite following Chivas et al. (1986). From these results, molar ratios of Mg/Ca $\times 10^{-2}$ and Sr/Ca $\times 10^{-3}$ were calculated. In total, 47 samples of recent ostracods and 18 samples of subfossil ostracods from 23 water bodies were used for the determination of Mg/Ca and Sr/Ca ratios.

The ostracod valves for oxygen isotope (δ^{18}O) and carbon isotope (δ^{13}C) analysis were prepared and cleaned in the same way as for the element analysis. To ensure enough material for isotope analysis (80–100 μg $CaCO_3$), 10–18 single valves of one species and sex from one waterbody were aggregated to create one sample. The aggregated samples were analysed by an isotope ratio gas mass-spectrometer (Finnigan MAT 251) directly coupled to an automated carbonate preparation device (Kiel II). The reproducibility, as determined by standard measurements over a 1-year period, is ±0.08‰ (1σ) for δ^{18}O and ±0.06‰ (1σ) for δ^{13}C. The values are expressed in delta per mil notation (δ, ‰) relative to the VPDB. In total, 46 samples of recent ostracods and 14 samples of subfossil ostracods from 23 water bodies were analysed for δ^{18}O and δ^{13}C stable isotopes.

Results

Physico-chemical characteristics of the ostracod habitats

During the summer period, when no ice covers the lakes and ponds, the water bodies are defined as polymictic. Because of the shallow water depth and the wind driven mixing, the water temperature is tightly correlated to the air temperature and ranged from 5.9°C to 15.3°C during the fieldwork period (Table 1). Dissolved oxygen concentrations were

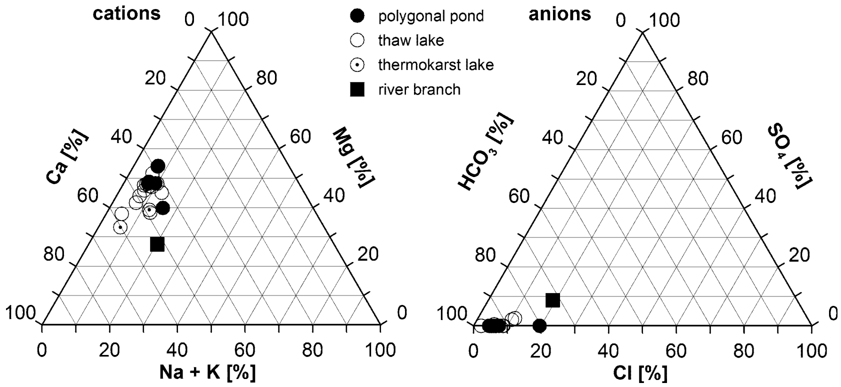


Fig. 3 Ionic composition in waters of lakes and ponds in the Lena River Delta

between 5.3 mg/l and 11.3 mg/l. The waters were oligotrophic (with phosphate values <0.1 ppm, below the detection limit), slightly acid to neutral (pH 6.5–7.6), and had alkalinities between 13.4 ppm and 148 ppm (Table 1). The electrical conductivity was very low and ranged from 27 µS/cm to 254 µS/cm. The waters were dominated by Mg–Ca or Ca–Mg, and HCO_3 (Fig. 3), but the ion content was generally low in all studied waters (Table 1). The molar Mg/Ca ratios ranged from 0.53 to 1.40 and the Sr/Ca ratios from 2.63×10^{-3} to 4.83×10^{-3} (Table 2).

The results of oxygen and hydrogen isotope analyses of the lake waters are presented in a δ^{18}O–δD plot (Fig. 4a) with respect to the Global Meteoric Water Line (GMWL), which correlates fresh surface waters on a global scale (Craig 1961). The studied lakes and ponds are mainly fed by precipitation waters. The isotope values of three rain water samples from the beginning of August 2002 and one sample of Lena River water are given in Fig. 4a. Whereas the local rain water samples and the river water are close to the GMWL, samples from the studied lakes and ponds are shifted below the GMWL. The deviation of the data from the GMWL reflects evaporation in the studied water bodies, as indicated by a slope of 5.56 ($R^2 = 0.96$; excluding the cut-off river branch) shown in Fig. 4a. The isotopic composition of one cut-off river branch (sample SAM-14) shows the influence of the Lena River with relatively lighter δ^{18}O (Fig. 4a).

The plot of δ^{18}O–δ^{13}C (Fig. 4b) shows no distinct differentiation in δ^{13}C between lakes and ponds situated on Samoylov and Kurungnakh islands. The waters varied between +0.1‰ and −14.4‰ for δ^{13}C. The δ^{13}C of the cut-off river branch observed on Samoylov Island (sample SAM-14) lies within this range. The wide range in δ^{13}C obviously reflects the influence of several biotic and abiotic factors on this parameter. The minimum value of −14.4‰ for δ^{13}C comes from an intrapolygon pond (sample SAM-13) on Kurungnakh Island. In this pond a pH-value of 6.5 was measured in the field. It had the lowest pH and consequently the lowest HCO_3 value of all the studied waters (Table 1). At constant temperatures a decrease in pH leads to a decrease in HCO_3 and therefore the composition of δ^{13}C in TDIC tends to become lighter (e.g., Clark and Fritz 1997).

The bottom sediments at the sites consisted primarily of minerogenic, sandy deposits with an organic cover. The main fraction of the studied sediments was composed of fine-grained sand (0.125–0.25 mm) or medium-grained sand (0.25–0.5 mm) (Fig. 5a). Furthermore, the sediments were characterised by C/N ratios from 9.8 to 17.9 (Fig. 5b), which are typical for polyhumic sediments with a low rate of decomposition (Hansen 1961) as is expected in high-latitude regions. The organic carbon content of the sediment varied between 0.7% and 17.3%. Highest values of more than 15% were found in sediments from two water bodies on Kurungnakh Island, which may be linked to higher organic carbon contents in the late Pleistocene Ice Complex deposits on the third Lena River terrace compared to those in the Holocene deposits on the first Lena River terrace (Schirrmeister et al. 2003).

Ostracod taxonomy and environmental ranges of their habitats

Among the 14 observed ostracod taxa, 11 taxa were identified down to the species level and two taxa to the genus level, whereas one taxon comprises

Table 2 Stable isotopes ($\delta^{18}O$, $\delta^{13}C$) and element ratios (Mg/Ca, Sr/Ca) of ostracod calcite and ambient water. The species identification follows the key: species (e.g., pedata → *F. pedata*), sex (f → female or m → male) and state (rec → recent or sub → subfossil)

Sample №	Species identification	Valve $\delta^{18}O$ ‰ PDB	Water $\delta^{18}O$ ‰ SMOW	Valve $\delta^{13}C$ ‰ PDB	Water $\delta^{13}C$ ‰ PDB	Valve Mg/Ca ($\times 10^{-2}$) Molar	Water Mg/Ca Molar	Valve Sr/Ca ($\times 10^{-3}$) Molar	Water Sr/Ca ($\times 10^{-3}$) Molar
SAM-01	pedata_f_rec	−13.55	−16.65	−3.75	−2.11	0.51	1.07	1.01	3.02
SAM-01	pedata_m_rec	−13.83	−16.65	−3.95	−2.11	0.37	1.07	1.06	3.02
SAM-12	pedata_f_rec	−15.92	−17.64	−8.44	−8.14	0.53	0.81	1.07	3.33
SAM-12	pedata_m_rec	−16.16	−17.64	−5.89	−8.14	0.48	0.81	0.98	3.33
SAM-12	pedata_f_sub	−15.47	−17.64	−7.15	−8.14	0.44	0.81	1.36	3.33
SAM-12	pedata_m_sub	−15.72	−17.64	−6.89	−8.14	0.20	0.81	1.52	3.33
SAM-13	pedata_f_rec	−11.38	−13.34	−11.01	−14.40	0.38	0.90	1.94	4.83
SAM-13	pedata_f_sub	−11.49	−13.34	−10.81	−14.40	0.21	0.90	1.67	4.83
SAM-13	pedata_m_sub	−10.81	−13.34	−11.10	−14.40	0.31	0.90	1.40	4.83
SAM-14	candida_f_rec	−17.69	−20.38	−7.74	−6.85	0.13	0.53	1.43	3.37
SAM-14	harmsworthi_f_rec	−18.46	−20.38	−6.47	−6.85	0.35	0.53	0.96	3.37
SAM-17	hyalina_f_rec	−14.21	−15.87	−2.76	−3.89	0.37	0.81	0.95	2.65
SAM-17	hyalina_m_rec	−14.12	−15.87	−2.85	−3.89	0.39	0.81	0.95	2.65
SAM-17	hyalina_f_sub	−13.97	−15.87	−2.98	−3.89	0.32	0.81	0.92	2.65
SAM-17	hyalina_m_sub	No data	−15.87	No data	−3.89	0.45	0.81	0.97	2.65
SAM-19	harmsworthi_f_rec	−15.14	−16.39	−3.40	−10.01	0.45	0.78	0.85	2.69
SAM-21	pedata_f_rec	−10.93	−13.29	−3.17	−0.22	0.41	1.39	0.93	3.09
SAM-21	pedata_m_rec	−11.00	−13.29	−3.50	−0.22	0.58	1.39	0.80	3.09
SAM-21	sanctipatricii_f_sub	No data	−13.29	No data	−0.22	1.00	1.39	1.18	3.09
SAM-23	pedata_f_rec	−12.66	−14.57	−2.38	−5.61	0.38	1.08	0.73	2.74
SAM-23	pedata_m_rec	−12.82	−14.57	−1.94	−5.61	0.57	1.08	0.69	2.74
SAM-23	species2_f_rec	−12.71	−14.57	−3.35	−5.61	1.08	1.08	0.89	2.74
SAM-23	species2_m_rec	No data	−14.57	No data	−5.61	0.62	1.08	0.85	2.74
SAM-23	harmsworthi_f_rec	−13.53	−14.57	−1.58	−5.61	0.87	1.08	0.69	2.74
SAM-23	harmsworthi_m_rec	−14.07	−14.57	−1.94	−5.61	0.84	1.08	0.75	2.74
SAM-24	pedata_f_rec	−10.94	−17.59	−4.19	−4.88	0.61	0.66	0.92	3.18
SAM-24	pedata_m_rec	−11.51	−17.59	−4.16	−4.88	0.46	0.66	0.92	3.18
SAM-24	pedata_f_sub	−10.80	−17.59	−4.19	−4.88	0.24	0.66	0.97	3.18
SAM-24	pedata_m_sub	−11.65	−17.59	−4.14	−4.88	0.47	0.66	0.96	3.18
SAM-25	pedata_f_rec	−13.31	−14.41	−3.72	−2.50	0.47	1.11	1.14	2.76
SAM-25	pedata_f_sub	−13.21	−14.41	−3.79	−2.50	0.39	1.11	0.73	2.76
SAM-25	pedata_m_sub	No data	−14.41	No data	−2.50	0.39	1.11	0.71	2.76
SAM-25	jakutica_f_sub	No data	−14.41	No data	−2.50	0.57	1.11	0.70	2.76
SAM-26	pedata_f_rec	−11.37	−13.46	−2.54	−9.12	0.29	1.15	1.06	3.00
SAM-26	pedata_m_rec	−11.78	−13.46	−3.44	−9.12	0.47	1.15	0.89	3.00
SAM-27	pedata_f_rec	−12.19	−14.11	−2.42	−1.39	0.41	1.11	0.77	2.94
SAM-27	pedata_m_rec	−12.45	−14.11	−1.84	−1.39	0.57	1.11	0.84	2.94
SAM-27	pedata_f_sub	−12.51	−14.11	−2.52	−1.39	0.51	1.11	0.80	2.94
SAM-28	pedata_f_rec	−11.71	−13.35	−2.48	0.05	0.37	0.98	0.85	2.77
SAM-28	pedata_m_rec	−11.52	−13.35	−2.84	0.05	0.52	0.98	1.10	2.77
SAM-29	pedata_f_rec	−12.56	−14.17	−4.16	−3.45	0.42	0.89	1.36	2.64

Table 2 continued

Sample №	Species identification	Valve δ^{18}O ‰ PDB	Water δ^{18}O ‰ SMOW	Valve δ^{13}C ‰ PDB	Water δ^{13}C ‰ PDB	Valve Mg/Ca ($\times 10^{-2}$) Molar	Water Mg/Ca Molar	Valve Sr/Ca ($\times 10^{-3}$) Molar	Water Sr/Ca ($\times 10^{-3}$) Molar
SAM-29	pedata_m_rec	−12.13	−14.17	−4.41	−3.45	0.46	0.89	1.23	2.64
SAM-30	pedata_f_rec	−10.69	−13.16	−1.82	−0.19	0.52	1.40	0.87	3.07
SAM-30	pedata_m_rec	−10.94	−13.16	−2.50	−0.19	0.67	1.40	0.92	3.07
SAM-32	pedata_f_rec	−11.80	−14.26	−2.55	−3.71	0.51	1.03	0.83	2.73
SAM-32	pedata_m_rec	−12.12	−14.26	−2.45	−3.71	0.32	1.03	0.81	2.73
SAM-32	pedata_f_sub	−11.55	−14.26	−2.11	−3.71	0.53	1.03	0.92	2.73
SAM-33	pedata_f_rec	−11.83	−14.01	−4.13	−1.13	0.54	1.06	1.13	2.75
SAM-33	pedata_m_rec	−12.20	−14.01	−4.89	−1.13	0.80	1.06	1.21	2.75
SAM-34	pedata_f_rec	−10.84	−13.80	−2.18	−1.95	0.39	1.24	0.81	2.78
SAM-34	pedata_m_rec	−11.76	−13.80	−2.53	−1.95	0.44	1.24	0.78	2.78
SAM-37	harmsworthi_f_rec	−13.32	−15.41	−3.89	−4.60	0.97	1.09	0.66	2.64
SAM-38	pedata_f_rec	−15.46	−17.39	−6.99	−3.49	0.77	0.55	0.92	3.22
SAM-38	pedata_m_rec	−15.17	−17.39	−6.75	−3.49	0.54	0.55	0.84	3.22
SAM-38	pedata_f_sub	−15.36	−17.39	−7.25	−3.49	0.25	0.55	0.84	3.22
SAM-38	pedata_m_sub	−14.95	−17.39	−6.82	−3.49	0.42	0.55	0.86	3.22
SAM-38	candida_f_rec	−14.97	−17.39	−6.91	−3.49	0.68	0.55	1.18	3.22
SAM-38	candida_f_sub	−14.74	−17.39	−6.57	−3.49	0.26	0.55	1.18	3.22
SAM-40	pedata_f_rec	−12.46	−15.37	−3.59	−4.00	0.59	1.07	1.04	2.96
SAM-40	pedata_m_rec	−12.19	−15.37	−4.15	−4.00	0.55	1.07	0.92	2.96
SAM-41	pedata_f_rec	−10.90	−13.81	−4.86	−6.89	0.21	1.06	0.74	2.63
SAM-41	pedata_m_rec	−11.75	−13.81	−4.97	−6.89	0.36	1.06	0.87	2.63
SAM-41	pedata_f_sub	−10.40	−13.81	−4.78	−6.89	0.49	1.06	0.82	2.63
SAM-44	pedata_f_rec	−11.42	−13.88	−6.26	−8.99	1.78	1.14	1.03	3.11
SAM-44	pedata_m_rec	−11.14	−13.88	−6.17	−8.99	0.32	1.14	0.95	3.11

indeterminate juvenile Candoninae. Here, we present the species from lakes and ponds that were used in describing the geochemical applications (Fig. 6). The ostracod assemblage in the shallow water bodies of the Lena River Delta consists partly of cosmopolitian species like *Candona candida* (O.F. Müller, 1776), *Fabaeformiscandona* cf. *hyalina* (Brady & Robertson, 1870), *Fabaeformiscandona* cf. *tricicatricosa* (Diebel & Pietrzeniuk, 1969), *Fabaeformiscandona protzi* (Hartwig, 1898), *Bradleystrandesia reticulata* (Zaddach, 1844), *Limnocytherina sanctipatricii* (Brady & Robertson, 1869), and *Cypria ophtalmica* (Jurine, 1820). The second group of ostracods found in lakes and ponds of the Lena River Delta are typical species for the Subarctic and Arctic of Siberia like *Candona muelleri jakutica* Pietrzeniuk, 1977, *Fabaeformiscandona harmsworthi* (Scott, 1899), *Fabaeformiscandona pedata* (Alm, 1914), and *Tonnacypris glacialis* (Sars, 1890). The two taxa *Fabaeformiscandona* sp. 1 and *Fabaeformiscandona* sp. 2 still remain in open nomenclature. Bisexual populations were found for all species except for *C. candida*, *C. ophtalmica*, and *T. glacialis*. The ostracod species frequency for the studied lakes and ponds is shown in Fig. 6. The species *C. muelleri jakutica*, *F. protzi*, *F.* sp. 2, and *B. reticulata* were mostly found in ponds and small lakes of polygonal genesis. A further dependence of species distribution on the water body type was not observed. The most common species in our study were *F. pedata* and *F. harmsworthi*. In Fig. 7 the ecological range is shown according to the environmental parameters

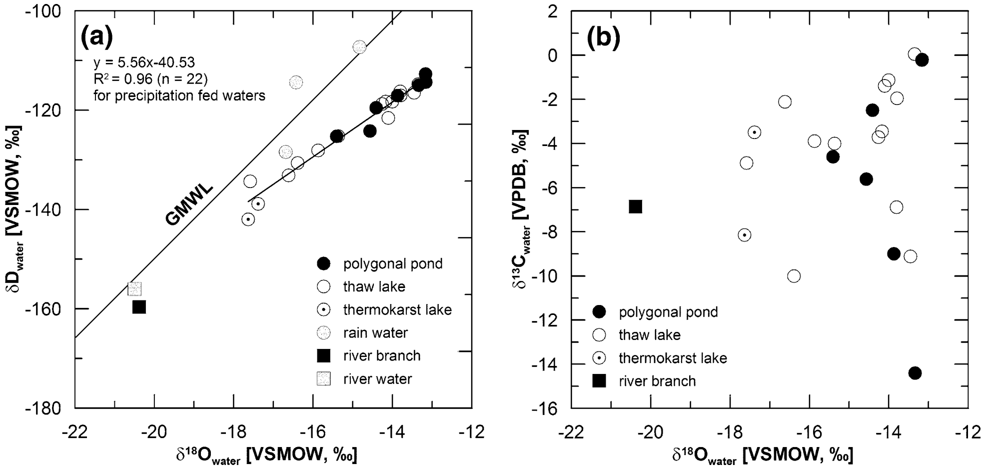


Fig. 4 Isotopic composition in natural waters in the Lena River Delta: (**a**) Plot of oxygen and hydrogen isotopes (δ^{18}O and δD) in lake water and precipitation in summer 2002 as well as in Lena River water; (**b**) Plot of oxygen and carbon isotopes (δ^{18}O and δ^{13}C) in lake water

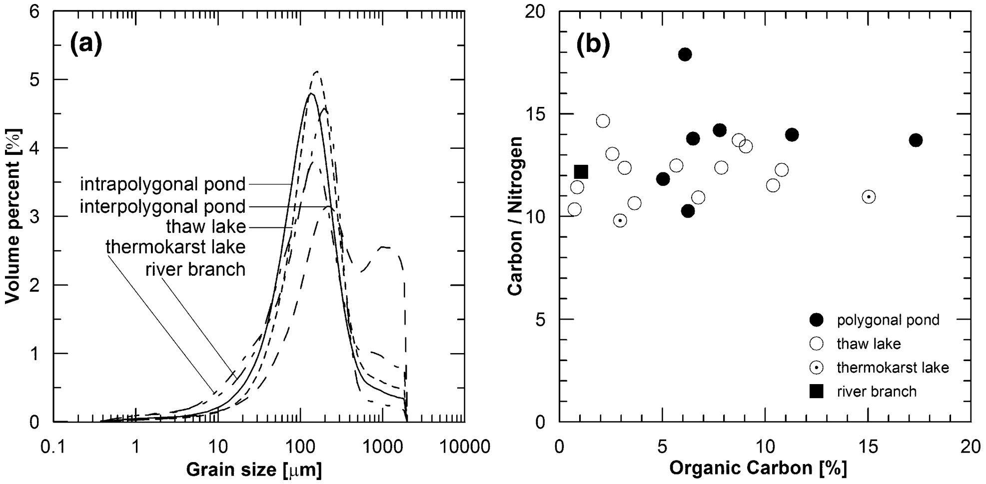


Fig. 5 Sedimentological properties of bottom sediments for the studied types of lakes and ponds: (**a**) Grain size distributions of examples for intrapolygon ponds (SAM-30); interpolygon ponds (SAM-37), thaw lakes (SAM-28), thermokarst lake shores (SAM-38) and river branches (SAM 14); (**b**) Plot of C/N ratios and organic carbon content in bottom sediments

at the sampling site, when the species occurred. Anyway, these data do not reflect the species tolerance as the only a limited gradient is covered by the studied waters.

In general, the gradient of the environmental parameters during the sampling time in August 2002 is quite low due to the very low ionic content of the studied water. The broadest ranges were found for the most common species in our study *F. pedata* and *F. harmsworthi*.

Ostracod geochemistry

The correlation between the $CaCO_3$ Nominal Shell Weight (NSW) and the measured valve weights is shown in Fig. 8. In general, the valve weights determined by micro-balance are higher than the calculated $CaCO_3$ NSW since ostracod valves consist of 80–90% $CaCO_3$ (Sohn 1958). Distinct or systematic differences between weights of recent and subfossil valves of any one species were not

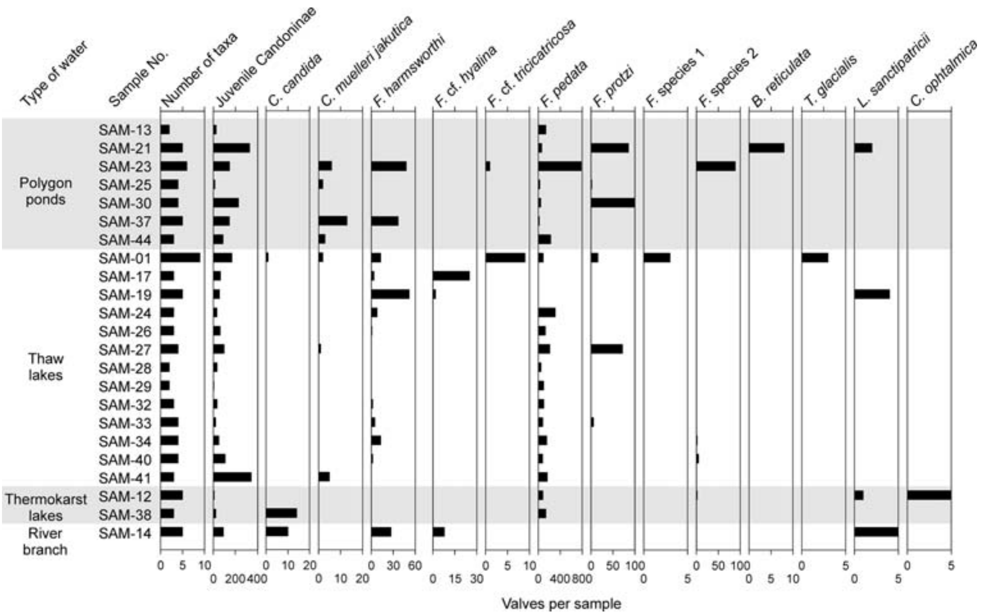


Fig. 6 Ostracod species frequency (in absolute numbers) in different types of lakes and ponds in the Lena River Delta. Note varying scales

observed. The reliability of the preparation method (soft body extraction) of the modern ostracods for geochemical analyses is therefore assumed.

The element ratios (Mg/Ca, Sr/Ca) in ostracod calcite are listed in Table 2. In Fig. 9 the relationship of element ratios in calcite for several species are shown. The highest variation in the Mg/Ca ratio is found in recent valves of *F. harmsworthi*, ranging from 0.35 to 0.97 × 10^{-2} (mean 0.7 × 10^{-2}; $n = 5$), whereas Mg/Ca ratios from the most common species in our dataset, *F. pedata*, range between 0.21 and 0.8 × 10^{-2} (mean 0.45 × 10^{-2}; $n = 49$). The Sr/Ca ratios for *F. harmsworthi* vary between 0.66 and 0.96 × 10^{-3} (mean 0.78 × 10^{-3}; $n = 5$), and for *F. pedata* between 0.69 and 1.94 × 10^{-3} (mean 1 × 10^{-3}; $n = 49$). Element ratios of subfossil valves lie within the ranges of recent valves (Fig. 9).

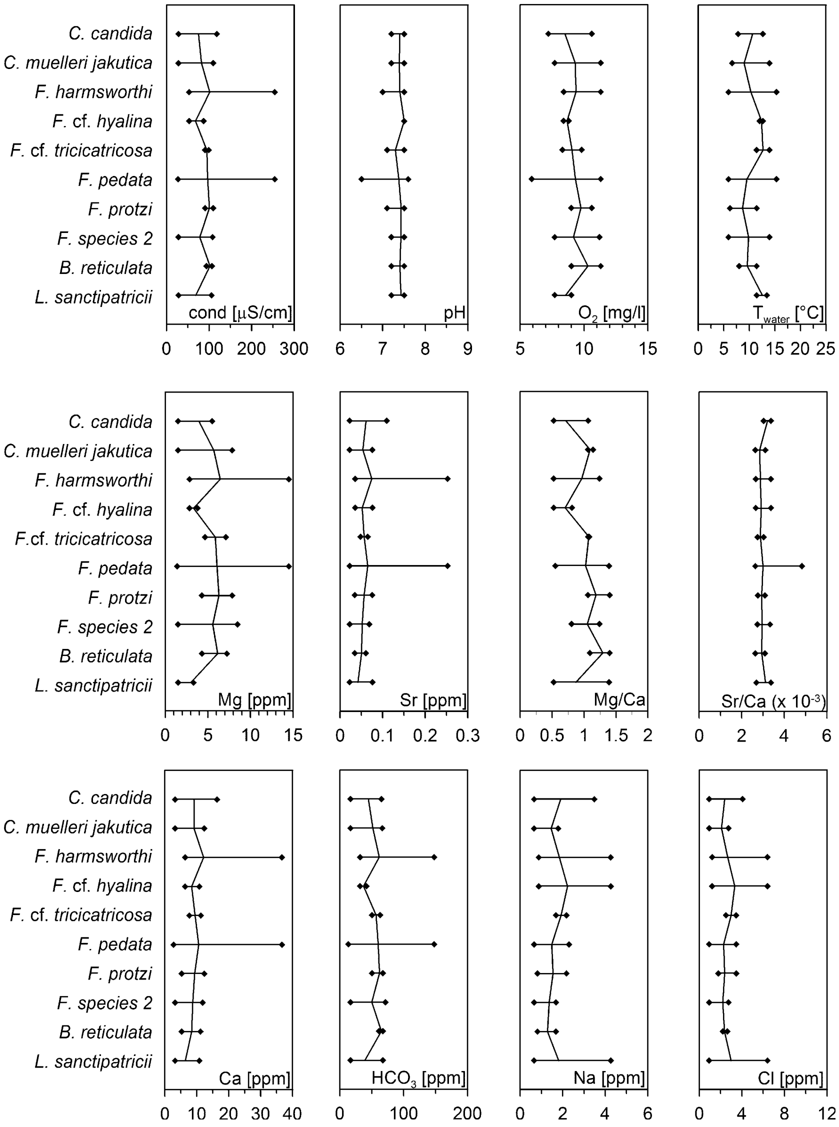
The plot of stable isotopes (δ^{18}O, δ^{13}C) in ostracod calcite reflects a differentiation with regard to the origin of water (Fig. 10). The isotopic composition of ostracod calcite from lakes and ponds on Samoylov Island fed by precipitation shows heavier isotopic composition, with values between −1.6‰ and −6.3‰ for δ^{13}C, and between −10.4‰ and −15.1‰ for δ^{18}O (Fig. 10). The waters on Kurungnakh Island are situated on the up to 40 m high third Lena River terrace and are not influenced by river water. The isotope record of ostracods from this location is lighter in comparison to the data from Samoylov Island in δ^{18}O or in both δ^{18}O and δ^{13}C (Fig. 10). The isotopic composition of ostracod calcite from one cut-off river branch on Samoylov Island (sample SAM-14) is characterised by relatively light values (-17.7 and -18.5‰ for δ^{18}O) as compared to those from precipitation-fed waters, whereas the δ^{13}C in ostracod calcite from the river branch does not differ (Fig. 10).

Discussion

Taxonomy and ecology of ostracods

The species *C. candida*, *F.* cf. *hyalina*, *F. protzi*, *B. reticulata*, *C. ophtalmica*, and *L. sanctipatricii* are known from mid-latitude regions to be broadly distributed and tolerant to a wide range of environmental factors, but with a preference or at least a tolerance for cooler water temperatures (Hiller 1972; Meisch 2000; Viehberg 2006). Their occurrence in Arctic environments shows their great ability for adaptation to extreme climatic conditions where the time for ontogenic cycles is very short. Semenova (2005) listed these species as commonly distributed

Fig. 7 Ranges of environmental parameters of ostracod habitats for most current taxa found in the studied shallow lakes and ponds. Horizontal lines connect the minimum and the maximum, and the vertical line is the mean. *C. candida* ($n = 3$), *C. muelleri jakutica* ($n = 7$), *F. harmsworthi* ($n = 12$), *F.* cf. *hyalina* ($n = 3$), *F. levanderi* ($n = 2$), *F. pedata* ($n = 21$), *F. protzi* ($n = 6$), *F.* sp. 2 ($n = 4$), *B. reticulata* (n = 3) and *L. sanctipatricii* ($n = 4$). Note varying scales



in East Siberia. Furthermore, *C. candida*, *F. hyalina*, and *B. reticulata* were found in thermokarst lakes in permafrost regions of Central Yakutia (Pietrzeniuk 1977). *C. candida* and *B. reticulata* were also described from Arctic environments on Greenland and in North Siberia (Alm 1914).

Species with preferences for (sub-) Arctic environments are *C. muelleri jakutica*, *F. harmsworthi*, *F. pedata*, and *T. glacialis*. The subspecies *C. muelleri jakutica* was first described by Pietrzeniuk (1977) from thermokarst lakes in Central Yakutia, whereas *F. harmsworthi* is commonly distributed in East Siberia (Semenova 2005) and known from Arctic environments on the Novaya Zemlya Archipelago and Franz Josef Land (Neale 1969). Another typical representative of Arctic freshwater ostracods is *F. pedata*, which was originally described as *Candona pedata* by Alm (1914). The genus *Fabaeformiscandona*, defined by Krstić (1972), did not originally include *F. pedata*, but the structure of the externo-distal seta (γ-seta) of the penultimate segment of the mandibular palp (which is smooth, not pulmose), and a carapace longer than 0.6 mm confirm the attribution of *F. pedata* to this genus. *T. glacialis* is known from East Siberia (Semenova 2005) and is considered to be endemic to most parts of the Arctic (Griffiths et al. 1998).

Some of the recently found species in the study area are known from Quaternary permafrost deposits

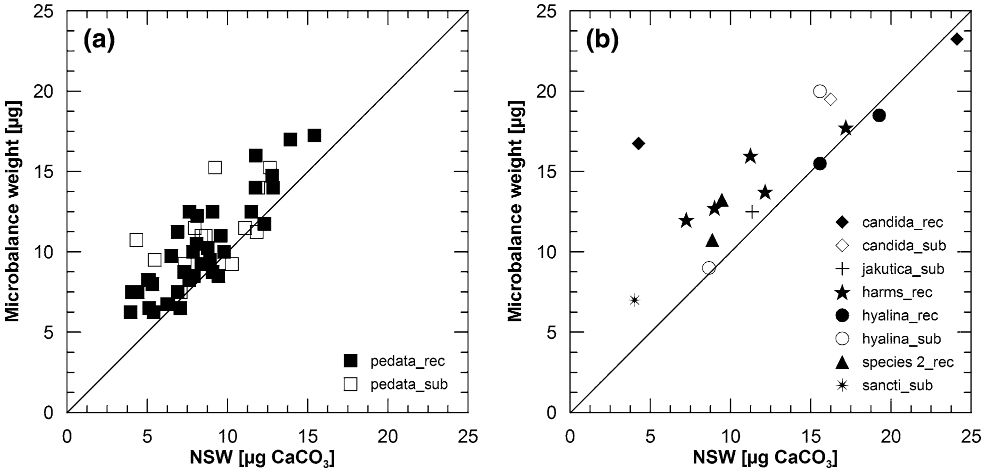


Fig. 8 Correlation of the calculated Nominal Shell Weight (NSW, Ca content by ICP-OES) and the measured weight (by micro-balance) of recent and subfossil single valves of: (**a**) *F. pedata*; (**b**) *C. candida*, *C. muelleri jakutica*, *F. harmsworthi*, *F.* cf. *hyalina*, *F.* sp. 2 and *L. sanctipatricii*. The species identification follows the key: species (e.g., *F. pedata* → pedata) and state (recent → rec or subfossil → sub)

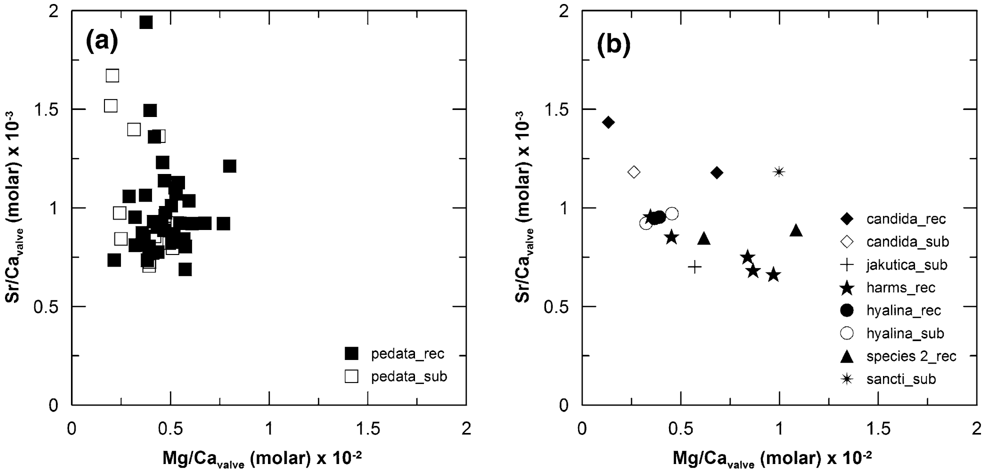


Fig. 9 Element ratios (Sr/Ca and Mg/Ca) of recent and subfossil single valves of: (**a**) *F. pedata*; (**b**) *C. candida*, *C. muelleri jakutica*, *F. harmsworthi*, *F.* cf. *hyalina*, *F.* sp. 2 and *L. sanctipatricii*. The species identification follows the key: species (e.g., *F. pedata* → pedata) and state (recent → rec or subfossil → sub)

in northeast Siberia, and therefore hold potential for aiding in regional palaeoenvironmental reconstructions. *C. candida*, *F. harmsworthi*, *F. pedata*, *L. sanctipatricii*, and *T. glacialis* are documented by Pietrzeniuk (unpublished data) in Late Pleistocene Ice Complex deposits in the Lena River Delta, and *C. muelleri jakutica*, *F. harmsworthi*, *F.* cf. *hyalina*, and *L. sanctipatricii* were collected in permafrost deposits on Bykovsky Peninsula to the east of our study sites (Wetterich et al. 2005).

Element ratios in ostracods and ambient waters

A relationship between the element content in ostracod calcite and water properties controlling the uptake of elements was established by studies in laboratory cultures and field collections of freshwater ostracods (e.g., Engstrom and Nelson 1991; Xia et al. 1997b; Wansard et al. 1998). This relationship is usually expressed as the partition coefficient *D*(M)

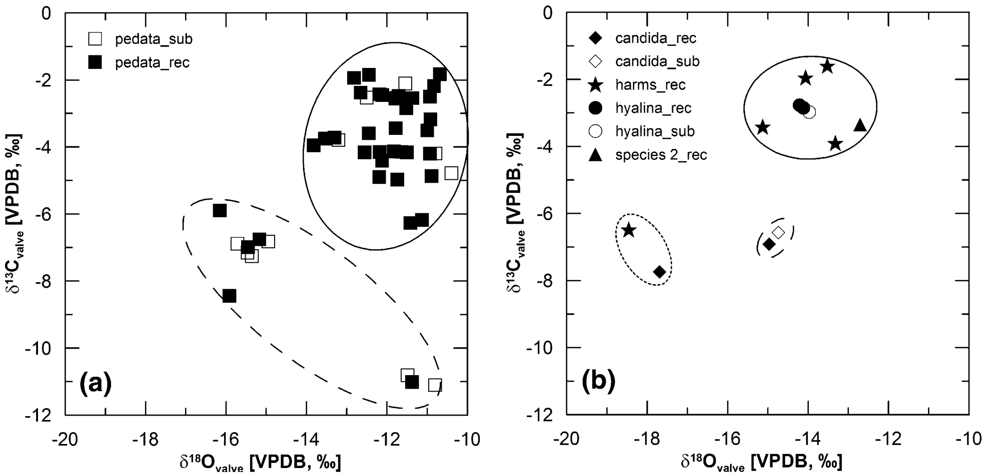


Fig. 10 Isotopic composition (δ^{18}O and δ^{13}C) in ostracod calcite for recent and subfossil single valves of: (**a**) *F. pedata*; (**b**) *C. candida*, *F. harmsworthi*, *F.* cf. *hyalina* and *F.* sp. 2. The species identification follows the key: species (e.g., *F. pedata* → pedata) and state (recent → rec or subfossil → sub). Full circles mark precipitation fed waters on Samoylov Island. Dashed circles mark precipitation fed waters on Kurungnakh Island. The dotted circle marks a river fed branch water on Samoylov Island

$$D(\mathrm{M}) = (\mathrm{M/Ca})_{\mathrm{valve}} \, / \, (\mathrm{M/Ca})_{\mathrm{water}} \qquad (1)$$

where M can either be Mg or Sr, and M/Ca ratios are molar ratios. This function reflects the environmental conditions of the host water (temperature, Mg/Ca, Sr/Ca) expressed as the Mg/Ca and Sr/Ca ratio of the valves of any one species at the time of calcification to the corresponding ratio in ambient water, and it is used for palaeoenvironmental reconstructions which are based on ostracod geochemical data (e.g., Chivas et al. 1986; Ricketts et al. 2001). Other factors which may also control the element composition of ostracod calcite, such as the biological effect of temperature dependent metabolic rates and ionic uptake also have to be taken into account (e.g., Dettmann et al. 2002). Presently the relationships between the element composition in host waters and ostracod calcite are not fully understood (Ito et al. 2003).

The magnesium uptake during valve calcification is a function of both the Mg/Ca ratio and the temperature of the water. Therefore, Mg/Ca ratios have the potential for being used as palaeothermometer (Chivas et al. 1986; Boomer et al. 2003). The observed Mg/Ca ratios in waters and ostracod calcite reach 1.39 in waters and 1.08 × 10^{-2} in valves. The lack of covariance between Sr and Mg (Fig. 9) can be explained by these low Mg/Ca ratios of the ambient waters and the narrow range of Mg/Ca in ostracod calcite (Xia et al. 1997b). The plot of Mg/Ca ratios for *F. pedata* versus water shows no covariance within the observed ranges due to the temperature-dependence of Mg partitioning and the very low Mg/Ca ratios in the ambient waters (Fig. 11a). For *F. harmsworthi*, which was investigated in much lower numbers, higher Mg/Ca ratios in waters seem to lead to higher ratios in valves (Fig. 11b). Partition coefficients for magnesium *D*(Mg) were not calculated because of their temperature-dependence, and therefore should be investigated in laboratory cultures under controlled temperatures only. Furthermore, water temperatures measured during fieldwork at the time of ostracod sampling were likely different from those at the time of calcification of the valves. Strontium uptake into ostracod calcite and consequently the resulting Sr/Ca ratios correlate with the Sr/Ca ratio (mostly depending on salinity) of the ambient water (Chivas et al. 1986; Engstrom and Nelson 1991). De Dekker et al. (1999) presented preliminary data that suggest the possibility of a small thermodependence of Sr uptake in ostracod calcite of *Cyprideis*. This relationship underscores the potential of using Sr/Ca ratios as a salinometer in palaeoenvironmental reconstructions based on the geochemistry of fossil ostracods (e.g., Chivas et al. 1986; Boomer et al. 2003). The Sr/Ca ratios observed in ostracod calcite in the Lena River Delta seem to be positively correlated with their ambient waters (Fig. 12). However, the variation in electrical

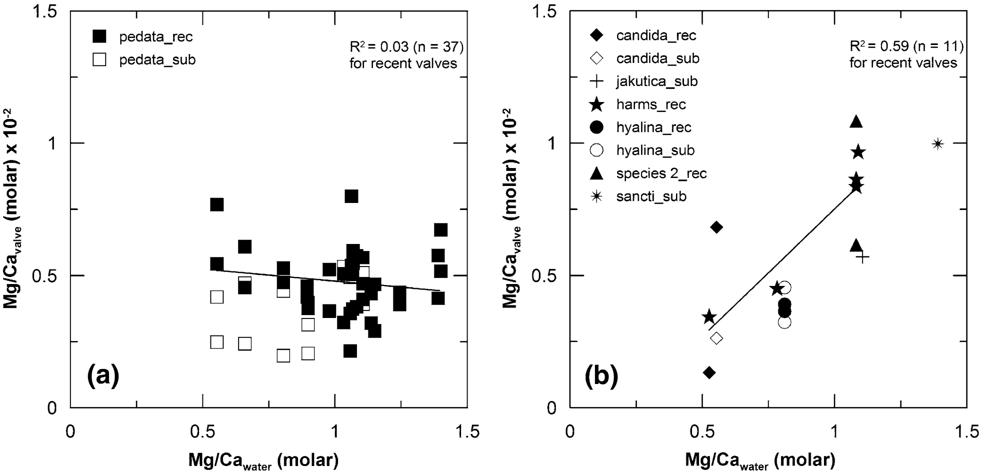


Fig. 11 Plot of Mg/Ca ratios in ostracod calcite and ambient waters for recent and subfossil single valves of: (**a**) *F. pedata*; (**b**) *C. candida, C. muelleri jakutica, F. harmsworthi, F.* cf. *hyalina, F.* sp. 2 and *L. sanctipatricii*. The species identification follows the key: species (e.g., *F. pedata* → pedata) and state (recent → rec or subfossil → sub)

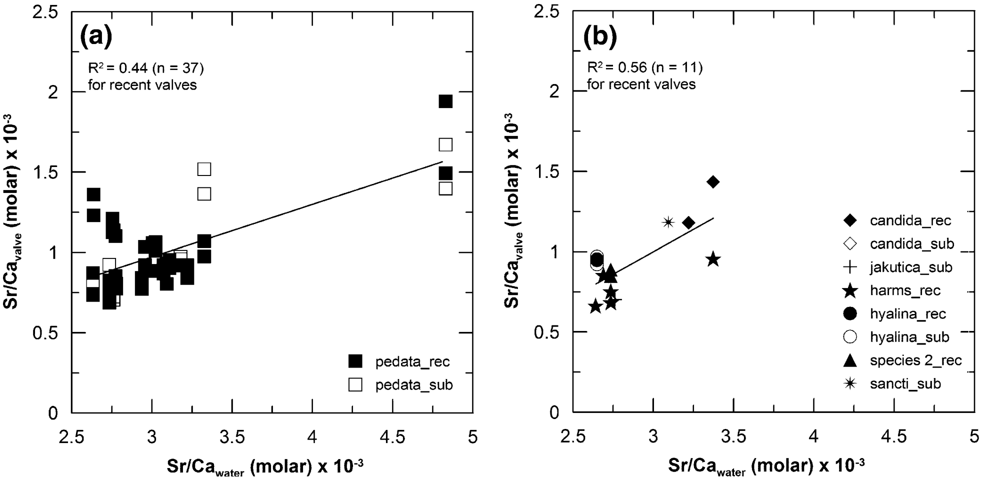


Fig. 12 Plot of Sr/Ca ratios in ostracod calcite and ambient waters for recent and subfossil single valves of: (**a**) *F. pedata*; (**b**) *C. candida, C. muelleri jakutica, F. harmsworthi, F. cf. hyalina, F.* sp. 2 and *L. sanctipatricii*. The species identification follows the key: species (e.g., *F. pedata* → pedata) and state (recent → rec or subfossil → sub)

conductivity in our data is too small to assume a strong correlation between changes in Sr/Ca ratio and electrical conductivity (salinity). The Sr/Ca ratio varies between 0.66 and 1.94 × 10^{-3} in ostracod calcite and between 2.63 and 4.83 × 10^{-3} in waters (Fig. 12). For the most common species in our study, *F. pedata*, average partition coefficients *D*(Sr) for living caught specimens were calculated according to equation (1) and were 0.33 ± 0.06 (1σ) for female adults ($n = 19$) and 0.32 ± 0.06 (1σ) for male adults ($n = 18$). Xia et al. (1997b) discussed a Mg dependence of *D*(Sr), where *D*(Sr) increases with Mg concentration because of the large physiological energy needed to exclude Mg (and Sr) during shell calcification in waters of high Mg/Ca. Following this assumption, in our data the calculated low partition coefficients *D*(Sr) correspond to the low Mg/Ca ratios of the ambient waters. *D*(Sr) values for other species were not calculated due to the low numbers of individuals.

Arctic ostracods are observed to produce only one population in the open water season (Semenova 2003), which lasts about 3 months in the Lena River Delta. That implies that the adults will hatch first in July. So the time lag between shell secretion and time of our sampling (in August) was quite short, while hydrochemical characteristics of the host waters were fairly stable. We therefore assume that shell chemistry at sampling time was similar to the one at secretion time. Nevertheless, it is clear that our results do not reflect the exact Mg/Ca and Sr/Ca ratios of host waters during shell secretion, because of laboratory manipulations, as shown in other experiments (Xia et al. 1997b). Furthermore, we emphasise strongly that the very low ionic content of the waters of the polygonal tundra does not allow a safe interpretation of the relationship between the element chemistry of the waters and the ostracod calcite. We refer to the fact that the observed variability does not allow us to calculate transfer functions.

In comparison with other studies on element chemistry in waters and ostracods (Engstrom and Nelson 1991; Xia et al. 1997b; Wansard et al. 1998, 1999; Wansard and Mezquita 2001), our data are likely to represent one endmember on the scale of hydrochemical information preserved in ostracods. It should be complemented by further investigations in (sub-) Arctic waters with higher ionic contents, where continental climatic conditions favour higher evaporation rates that alter the hydrochemical setting of fresh waters.

Stable isotopes in ostracods and ambient waters

Like element ratios, the isotopic composition of ostracod calcite ($\delta^{18}O$, $\delta^{13}C$) can be linked to the composition of the ambient water at the time of valve formation. (e.g., Chivas et al. 1993; Xia et al. 1997a, b; von Grafenstein et al. 1999). Therefore, $\delta^{18}O$ and $\delta^{13}C$ records of ostracod calcite provide a highly localised and temporally restricted reflection of the isotopic composition of water and TDIC, making them useful proxies in palaeolimnology (Holmes 1996). The isotopic composition of ostracod calcite shows both interspecific and intraspecific variations, mainly because of species-dependent metabolic effects on isotope fractionation, the timing of shell calcification in different seasons and at different temperatures, and species-dependent preferences for different microhabitats (e.g., Heaton et al. 1995; von Grafenstein et al. 1999). The $\delta^{18}O$ of ostracod carbonates is used as a proxy for temperature and isotopic composition of the water from which they precipitated (e.g., Chivas et al. 1993; Xia et al. 1997a; von Grafenstein et al. 1999). Factors affecting the oxygen isotope composition of lake water are the isotopic composition of precipitation, drainage basin hydrology, groundwater input, the precipitation/evaporation ratio, the residence time of water, the size of the waterbody, as well as the hydrochemical properties and temperature of the lake water (e.g., Kelts and Talbot 1990; Schwalb 2003; Leng and Marshall 2004).

Water bodies mostly fed by precipitation are common on the Lena River terraces above the floodplain and are generally not deep enough for extensive melting of the permafrost. Therefore, the influence of river water and meltwater from the frozen ground is of minor importance for most of the waterbodies. The main water source is summer precipitation. The $\delta^{18}O$ of these waterbodies ranged between −13.1‰ and −17.6‰ $\delta^{18}O$ (mean −15.1‰; $n = 23$) with a slope of 5.56 (Fig. 4a). According to Meyer et al. (2002), rain water samples from the Bykovsky Peninsula ranged between −11.4‰ and −17.0‰ $\delta^{18}O$ (mean −14.8‰; $n = 10$) with a slope of 6.77 and lie nearer to the GMWL than the data from lakes and ponds, which confirms the noticeable influence of evaporation on the $\delta^{18}O$ signal in waterbodies fed by precipitation. In general, the smaller polygonal ponds show a stronger deviation from the GMWL than thaw lakes, which in turn deviate more than thermokarst lakes (Fig. 4a); this is obviously caused by lower evaporation rates per volume in deeper waters. Furthermore, deeper thaw lakes and thermokarst lakes reflect a mixed isotopic signal of precipitation waters and meltwater from the underlying ground ice. The latter is known to have a mean $\delta^{18}O$ isotopic ratio of −25‰ in Holocene ground ice and −30‰ in late Pleistocene ground ice (Meyer et al. 2002). The isotopic signal of one cut-off river branch on Samoylov Island (sample SAM-14, −20.4‰ for $\delta^{18}O$ and −159.6‰ for δD) corresponds to the relatively light isotopic composition of the river water (−20.5‰ for $\delta^{18}O$ and −156‰ for δD;

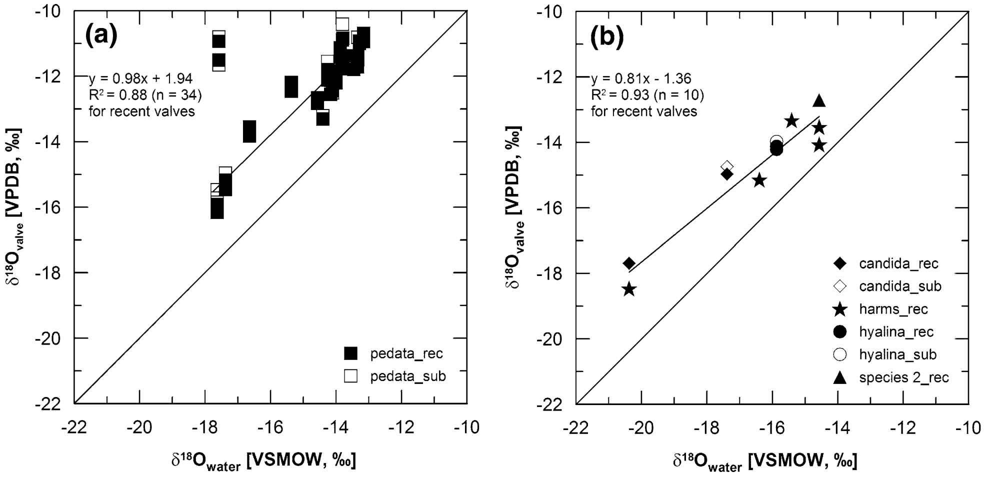


Fig. 13 Plot of δ^{18}O in ostracod calcite and ambient waters for recent and subfossil single valves of: (**a**) *F. pedata*; (**b**) *C. candida*, *F. harmsworthi*, *F.* cf. *hyalina* and *F.* sp. 2. The species identification follows the key: species (e.g., *F. pedata* → pedata) and state (recent → rec or subfossil → sub)

Schirrmeister et al. 2003). Obviously, this waterbody was flooded by the Lena River during spring flooding.

A strong covariance is shown between δ^{18}O in ostracod calcite of *F. pedata* (Fig. 13a) and of other species (Fig. 13b) with the water in which the valves were formed. The data show a near 1:1 relationship of δ^{18}O with a systematic shift to heavier values in ostracod calcite than in the respective ambient waters. According to Xia et al. (1997a), this shift (Δ) reflects the dependence between the δ^{18}O of ostracod calcite and of water with respect to water temperature, and is defined as:

$$\Delta = \delta^{18}\mathrm{O}_{\mathrm{valve}} - \delta^{18}\mathrm{O}_{\mathrm{water}}[\text{‰}] \qquad (2)$$

The shift varies between $\Delta_{\min} = +1.1$‰ and $\Delta_{\max} = +3.2$‰ (excluding two outliers), with a mean of + 2.2‰ ± 0.5 (1σ) for living caught female and male specimens of *F. pedata* (Fig. 13a). Two outliers with shifts of −6.1‰ and −6.7‰ were observed in valves of living female and male *F. pedata* caught in one thaw lake (SAM-24) on Samoylov Island. This thaw lake drained into the river, had a water depth less than one meter, and the highest electrical conductivity and ionic contents of all studied waters (Table 1). Probably because of the very shallow water in this lake, evaporation had a stronger influence than in other waterbodies and led to the unusual isotopic composition. These samples were therefore not included in further interpretation.

The parameters ($\delta^{18}\mathrm{O}_{\mathrm{valve}}$, $\delta^{18}\mathrm{O}_{\mathrm{water}}$) were measured in duplicate with high-precision of about ±0.1‰ (1σ) and therefore measurement errors can be ruled out. The isotopic composition of the water is more negative than that of the valves. However, the water was sampled in summer after the ostracod calcite had formed and $\delta^{18}\mathrm{O}_{\mathrm{water}}$ should consequently rather be heavier (more summer precipitation, more evaporation) than the measured one. This leads to the assumption that the systematic deviation of about 2‰ is not related to $\delta^{18}\mathrm{O}_{\mathrm{water}}$. The shift of about 2‰ (Fig. 13a) between $\delta^{18}\mathrm{O}_{\mathrm{water}}$ and $\delta^{18}\mathrm{O}_{\mathrm{valve}}$ includes most likely metabolic (vital) and temperature effects which cause this systematic deviation. A metabolic or vital offset as compared to inorganic calcite in equilibrium was quantified to about 1.4‰ for *F. pedata* (Fig. 14a). Vital offsets were already proposed by von Grafenstein et al. (1999), where a temperature-independent deviation (vital offset) of $\Delta = +2.2$‰ ± 0.15 (1σ) was inferred for several species of Candoninae. Other studies already showed the influence of vital effects on the isotopic composition of ostracod calcite of modern *F. rawsoni* (Xia et al. 1997a) and of *C. candida* (Keatings et al. 2002).

Our dataset is based on sampling of ostracods from shallow waterbodies with a high daily temperature range. Hence, it is difficult to relate our temperature data directly to $\delta^{18}\mathrm{O}_{\mathrm{water}}$ and $\delta^{18}\mathrm{O}_{\mathrm{valve}}$. Therefore, we correlated the minima and maxima of temperatures

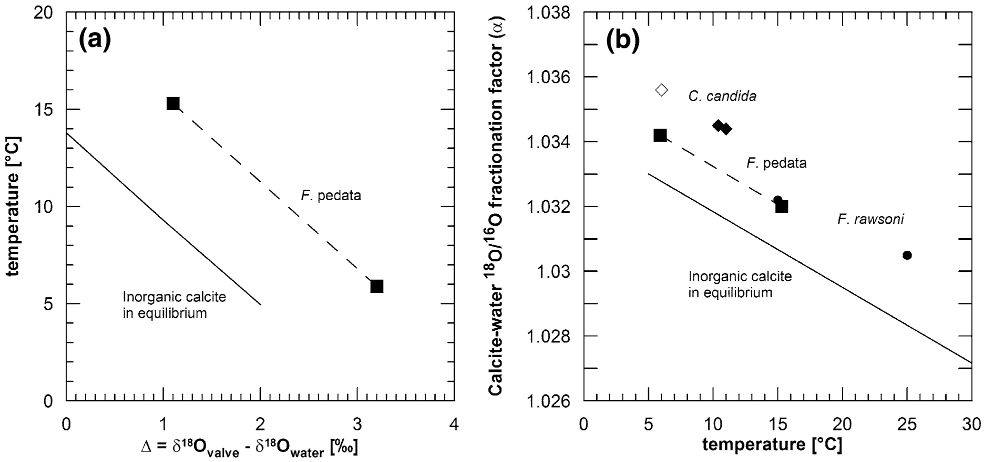


Fig. 14 Temperature-dependent δ^{18}O fractionation in *Fabaeformiscandona pedata* expressed as: (**a**) variations in the calcite-water oxygen isotope fractionation shift Δ (T [°C] = 20.22–4.48($\delta^{18}O_{valve}$–$\delta^{18}O_{water}$)) in comparison with that of inorganic carbonates (T [°C] = 13.8–4.58($\delta^{18}O_{valve}$–$\delta^{18}O_{water}$) + 0.08($\delta^{18}O_{valve}$–$\delta^{18}O_{water}$)^2) by Kim and O'Neil (1997), re-expressed by Leng and Marshall (2004); (**b**) variations in the calcite-water oxygen isotope fractionation factor α for ostracods of the genera *Candona* and *Fabaeformiscandona*: closed dots for *F. rawsoni* (Xia et al. 1997a); open diamond for *C. candida* (von Grafenstein et al. 1999); closed diamonds for *C. candida* (Keatings et al. 2002) and closed squares for *F. pedata* (this study) in comparison with data of equilibrium fractionation in inorganic calcite at 5 mM Ca^{2+} solution (Kim and O'Neil 1997) given as a solid line. The dashed line reflects the slope in the data of this study

(T_{min}, T_{max}) and of shifts (Δ_{min}, Δ_{max}) in our dataset (Fig. 14). Water temperatures and $\delta^{18}O_{water}$ measured at the time of ostracod sampling may have been different from those at the time of valve calcification. Nevertheless, we assume that field data reflect potential temperature variations between about 6 and 15°C over the summer period. Due to the lack of continuous water temperature measurements over the ice free period, we use field observations as well as air and soil temperature measurements from Samoylov Island in summer 2002 (Wille et al. 2003) to support the relevance of water temperature variations. The temperature regime of the studied shallow polymictic ponds and lakes is controlled by air and soil temperatures and should range between them. Hourly mean temperatures in July 2002 varied from 0°C to 25°C in the air and from 3°C to 10°C in the soil. In August 2002, the temperature variation ranged from 1°C to 24°C in the air and from 4°C to 10°C in the soil (J. Boike, AWI Potsdam, unpublished data). The studied shallow water bodies frozen down to the bottom in winter usually start thawing in the middle of June and refreeze in the middle of September (G. Stoof, AWI Potsdam, pers. comm.). The ostracods from these habitats most likely start their ontogeny not earlier than the waters are free of ice at the end of June when water temperature should be distinctly above 0°C. After this ostracods certainly need some time to reach adulthood. From these presumptions we conclude that a range of near bottom water temperature between about 6°C and 15°C which was measured during ostracod sampling seems to be quite realistic.

The temperature-dependence of δ^{18}O fractionation is reflected by the variation of the shift within a species, where increased temperatures correspond to smaller shifts (e.g., Leng and Marshall 2004). Xia et al. (1997a) showed a clear and consistent temperature-dependence of oxygen isotope fractionation during biological calcification by the species *Fabaeformiscandona rawsoni* (TRESSLER, 1957) in culture experiments at 15°C and 25°C. The 15°C cultures were about 2‰ heavier than the 25°C cultures, which confirms with the expected deviation of about 2‰ for a temperature difference of 10°C in inorganic carbonates (Xia et al. 1997a). The observed variation in the shift for modern *F. pedata* (between Δ_{min} = + 1.1‰ and Δ_{max} = + 3.2‰) over a temperature range of 9.4°C (T_{min} = 5.9°C and T_{max} = 15.3°C) during the fieldwork in summer 2002 can be explained by different temperatures of the water at the time of calcification (Fig. 14a). The equation relating δ^{18}O of ostracod calcite to temperature is: T [°C] = 20.22–4.48($\delta^{18}O_{valve}$–$\delta^{18}O_{water}$), defined according to the standard palaeotemperature scale by Epstein et al. (1953).

From the two samples where maximal and minimal shifts are found, the calcite-water oxygen isotope fractionation factors (α) are calculated based on the definition:

$$\alpha = \delta^{18}O_{valve} + 1{,}000/\delta^{18}O_{water} + 1{,}000 \qquad (3)$$

The results are α = 1.0342 at 5.9°C and α = 1.0320 at 15.3°C ($\delta^{18}O_{valve}$ and $\delta^{18}O_{water}$, relative to VSMOW). The temperature-dependence between ostracod calcite and ambient water, expressed as oxygen isotope fractionation factors (α), is given in Fig. 14b, along with results for inorganic calcite (Kim and O'Neil 1997) and for ostracod calcite from other studies (Xia et al. 1997a; von Grafenstein et al. 1999; Keatings et al. 2002) which dealt with species of the genera *Candona* and *Fabaeformiscandona*. The oxygen isotope fractionation factors for inorganic calcite in equilibrium are given as solid line for a 5 mM Ca^{2+} solution which was assumed as most representative for equilibrium fractionation (Kim and O'Neil 1997). The other data plotted in Fig. 14b derive from an in vivo study of *C. candida* in small shallow ponds in southern England (α = 1.0345 at 11°C; Keatings et al. 2002), and otherwise from an in vivo study of the same species in a large, deep lake in southern Germany (α = 1.0356 at 6°C, calculated by Keatings et al. 2002, based on data of von Grafenstein et al. 1999). Furthermore, data of an in vitro study of is shown for *F. rawsoni* (α = 1.0322 at 15°C and α = 1.0305 at 25°C; Xia et al. 1997a). The authors stated that *F. rawsoni* incorporate relatively more ^{18}O relative to inorganic calcite at 25°C than at 15°C, probably due to slower or less robust calcification, and to stress at the lower temperature. Therefore, the *F. rawsoni* data are closer to equilibrium fractionation line of inorganic calcite and reflect lower fractionation at a given temperature (Xia et al. 1997a).

Even though the comparison of data from different environments and laboratory experiments is highly speculative, the results are instructive since presently there are only a few studies on modern ostracods of different taxa and some aspects are worth discussing. Since the slope of temperature-dependence of the *F. pedata* data is defined by just two points (Fig. 14b), and our species is different, we only assume a general accordance with the results of the studies cited above. The effect of lower calcification probably due to temperature-stress conditions as described by Xia et al. (1997a) is not seen in our data and most likely due to coldwater adaptation by the Arctic species *F. pedata*. The slope between the two data points of *F. pedata* is parallel to that of equilibrium fractionation of inorganic calcite, but systematically shifted. Both data points are arranged between that of *Candona* species and inorganic calcite. Possible explanations of lower fractionation in *F. pedata* can be attributed to species-dependent vital effects which result in different fractionation factors. Up to now the mechanisms of ostracod calcification still remain unclear (e.g., Keatings et al. 2002). Furthermore, the different fractionation can be also caused by habitat-specific characteristics of Arctic environments, which are not fully understood up to now.

In summary, the variation in $\delta^{18}O$ ostracod calcite corresponds to the isotopic composition of ambient waters, which is affected by the general climatic setting of the region, the water source feeding the waterbody (precipitation or river water), and the influence of meltwater from the frozen ground. For the $\delta^{18}O$ of ostracods adapted to cold environments such as *F. pedata*, a temperature-dependence is reflected in the variations of calcite-water oxygen isotope fractionation factor (α) and shift (Δ). Additionally, the deviation to equilibrium fractionation is most likely influenced by metabolic (vital) effects. These vital effects are also seen in the deviation of the $\delta^{18}O$ signal in ostracod calcite of other species.

The $\delta^{13}C$ of carbonates is not very influenced by temperature variations but is rather understood to be a response to changes in the isotopic ratio of the total dissolved inorganic carbon (TDIC) from which the carbonates precipitated (Holmes 1996). Commonly, changes in $\delta^{13}C$ are attributed to changes in carbon and productivity within a lake (e.g., Schwalb 2003; Leng and Marshall 2004). Rates of exchange of CO_2 with the atmosphere, photosythesis/respiration of aquatic plants, organic decay, and bacterial processes are the main controlling factors for the $\delta^{13}C$ of TDIC (e.g., von Grafenstein et al. 1999; Schwalb 2003; Leng and Marshall 2004). In Fig. 15 the relationship between $\delta^{13}C$ in ostracod calcite of different species and in water is shown. The $\delta^{13}C$ of the waters ranges

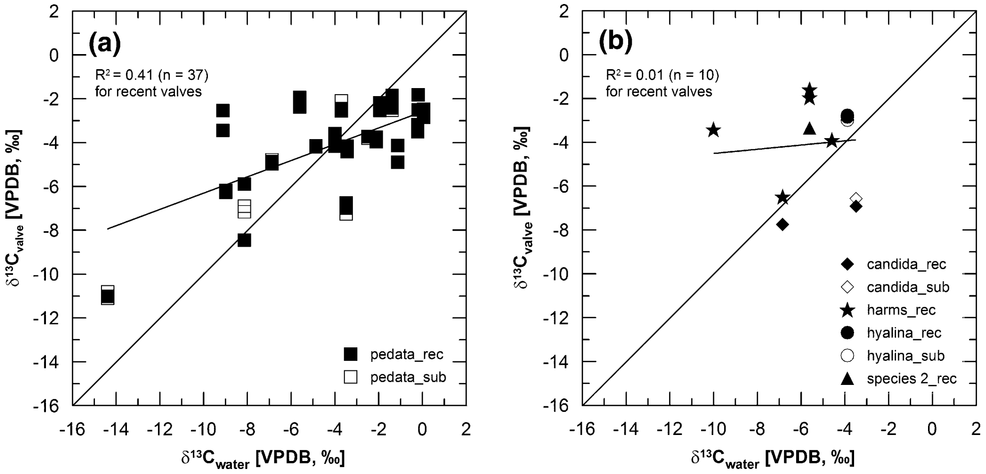


Fig. 15 Plot of δ^{13}C in ostracod calcite and ambient waters for recent and subfossil single valves of: (**a**) *F. pedata*; (**b**) *C. candida*, *F. harmsworthi*, *F.* cf. *hyalina* and *F.* sp. 2. The species identification follows the key: species (e.g. *F. pedata* → pedata) and state (recent → rec or subfossil → sub)

between +0.1‰ and −14.4‰, and waters fed by precipitation cannot be distinguished from that in the old branch. The considerable scatter in δ^{13}C indicates the influence of complex abiotic and biotic effects on δ^{13}C fractionation as is expected in natural lacustrine systems (Leng and Marshall 2004). The most probable explanation for the observed δ^{13}C variation at the studied shallow ponds and lakes is the daily variation of water plant photosynthesis and seasonal variations during the summer.

Conclusions

For the first time, this study deals with the geochemical properties of modern freshwater ostracods from northeast Siberia. The geochemical record in ostracod calcite (Mg/Ca, Sr/Ca; δ^{18}O, δ^{13}C) was investigated in comparison with data from ambient waters. Over the years, several studies on this branch of ostracodology have been undertaken, but even more recent studies do not fully explain the relationships between water and shell chemistry (e.g., Wansard et al. 1998, 1999; Griffiths and Holmes 2000; Boomer et al. 2003). Our study was the first attempt to expand knowledge about the geochemistry of freshwater ostracods in Arctic regions. The results show the potential of Arctic freshwater ostracods, which are often preserved in Quaternary permafrost deposits, as geochemical proxies for regional reconstructions of palaeoenvironments. The following conclusions can be drawn from our study:

(1) The ostracod assemblages in the Lena River Delta include typical Arctic species like *Candona muelleri jakutica*, *Fabaeformiscandona harmsworthi*, *Fabaeformiscandona pedata*, and *Tonnacypris glacialis*, but also cosmopolitan species like *Candona candida*, *Fabaeformiscandona* cf. *hyalina*, *Fabaeformiscandona* cf. *tricicatricosa*, *Fabaeformiscandona protzi*, *Bradleystrandesia reticulata*, *Cypria ophtalmica*, and *Limnocytherina sanctipatricii* were found. The most common species are *F. pedata* and *F. harmsworthi*.

(2) Due to the very low ionic content of the waters of the Arctic tundra, the observed element ratios of ostracod calcite (Mg/Ca, Sr/Ca) in recent valves of female and male specimens of different species range in very narrow arrays. Therefore, our data represent one endmember on a scale of hydrochemical information preserved in ostracods.

(3) No distinct differentiations are observed in Mg/Ca, Sr/Ca, δ^{18}O, and δ^{13}C between female and male or recent and subfossil valves, either within one species or between different species.

(4) Average partition coefficients D(Sr) for living caught specimens of *F. pedata* were calculated, with 0.33 ± 0.06 (1σ) for female adults

(n = 19), and 0.32 ± 0.06 (1σ) for male adults (n = 18).

(5) The δ^{18}O data of the waters, and consequently of ostracod calcite, indicate the water source in the watershed is either precipitation or the Lena River.

(6) A near 1:1 relationship of δ^{18}O in waters and valves was found, with a mean shift of Δ_{mean} = 2.2‰ ± 0.5 (1σ) to heavier values for calcite of modern *F. pedata* (n = 34) as compared to ambient waters. The shift is not dependent on $\delta^{18}O_{water}$, and caused by vital and temperature effects. Temperature-dependence is reflected in the variations of Δ (between Δ_{min} = +1.1‰ and Δ_{max} = +3.2‰). A vital effect as compared to inorganic calcite in equilibrium was quantified with 1.4‰ for *F. pedata*.

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