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Lake Imandra depression in the Late Glacial and early Holocene (Kola Peninsula, north-western Russia)

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Abstract. The paper summarizes the evidence of litho-, biostratigraphy and ^{14}C dating of sedimentary sequences studied in natural outcrops and bottom deposits in small lakes, as well as data on coastal morphology in the depressions of Ekostrovskaya and Babinskaya Imandra, the southern sub-basins of Lake Imandra. Lithological, ^{14}C and diatom data suggest that the brackish-water reservoir followed by the fresh-water one existed in the Ekostrovskaya Imandra depression during the Younger Dryas chronozone prior to 11,400cal. yr BP. The Fennoscandian Ice Sheet margin is assumed to have been located in the Lake Imandra basin, covering western Babinskaya Imandra earlier than *c.* 10,250 cal. yr BP. The early Holocene *c.* 11,400–8,500 cal. yr BP was marked by a significant westward retreat of the ice margin in the western Lake Imandra depression and adjacent areas, and an extensive fresh-water pra-Imandra Lake basin was formed there. At the end of the Preboreal, earlier than *c.* 9,210–8,500 cal. yr BP, the pra-Imandra Lake coastline was at least 16–18 m higher than the modern one, as can be assumed according to coastal morphology and lithostratigraphical data. The coastline of that reservoir changed, water square slightly reduced, and isolated small lakes emerged on coasts during the early Holocene.

Keywords: *lithostratigraphy; diatoms; bottom sediments; natural outcrops; coastal morphology; periglacial fresh-water reservoir; marine inundation*

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INTRODUCTION

Lake Imandra is the largest water body in the Murmansk region. Within its catchment area, more than 200 thousand people live and a number of industrial enterprises operate, including a nuclear power plant with a direct-flow cooling system. Along the lake shores, there is a federal motor road and a railway. For more than half a century, Lake Imandra with lakes and streams in its catchment area have been used as a source for technical and potable water, as well as for purposes of recreation, tourism and fishing. Hence, not only the ecosystem evolution and economic resources of Lake Imandra, but also the geological history and palaeoenvironment are a crucial matter of study. In the second half of the past century, research was focused on the tectonics and geomorphology of the central Kola region and geodynamic and paleogeographic interpretation of geological, biostratigraphical and geomorphological data (Armand *et al*. 1964, 1969; Fedorov 1965; Armand, Samsonova 1969; Yakovlev 1974; Kagan *et al*. 1980; Koshechkin *et al*. 1980; etc.). Thereafter, the dynamics of the Lake Imandra compound basin, tectonic events and their causes in the Lake Imandra depression and adjacent areas, deglaciation and glacioisostasy, changes in the hydrological regime and stratigraphy of bottom sediments have been the most discussed issues (Hättestrand, Clark 2006; Nikolaeva *et al*. 2015, 2016, 2018; Tolstobrova *et al*. 2016; Korsakova *et al*. 2019; Lenz *et al*. 2020). Nowadays, the earliest evolution of Lake Imandra associated with deglaciation and development of Lake pra-Imandra at initial stages in the Late Glacial and early Holocene is the most debated issue that requires examination. The current study summarizes data on coastal topography, depositional sequences and bottom sediments from small lake basins in coastal areas of Lake Imandra and aims at palaeoenvironmental reconstructions of Lake pra-Imandra in the Younger Dryas and early Holocene.

STUDY AREA

Lake Imandra (127 m above sea level (a.s.l.)) is the largest lake in the eastern Fennoscandian crystalline shield. Its eastern coast relates to the Kola Peninsula, while the western coastal areas occur in the continental part of Fennoscandia. Today, the lake consists of three sub-basins connected by narrow straits (salmas). In the north, Bolshaya (Khibinian) Imandra has a submeridional orientation and merges with Ekostrovskaya Imandra via the Ekostrovsky Strait. In the south, Ekostrovskaya Imandra and Babinskaya Imandra are oriented in the sublatitudional direction and connected via the Shirokaya Salma Strait (Fig. 1). The southern part of the Lake Imandra basin is the main focus of the current study.

The topography of the studied coastal area ranges from 127 to 180–190 m a.s.l. in the Bolshaya Imandra basin and to 130–190 m a.s.l. in the Ekostrovskaya and Babinskaya Imandra depressions. The bedrock of the area is composed of the Late Archaean and Early Proterozoic gneisses, plagiogneisses, migmatites, gneiss-granites, and amphibolites. Besides, alkaline-ultrabasic intrusions (Niva, Afrikanda, Ozernaya Varaka and Lesnaya Varaka) and the Khibiny alkaline massif occur in the south-east and north-east, respectively (Fig. 2). The Quaternary deposits mainly consist of glacial and glaciofluvial or glaciolimnic sediments, which produce till covers on the bedrock, local hummocky and end moraines, numerous drumlins and eskers, deltas and other glaciofluvial formations. It is noteworthy that eskers elongate in the north-west-south-east direction mostly, and drumlins are oriented from west to east (Punkari 1995; Stroeven *et al*. 2016). Some local depressions show Holocene lacustrine sediments in the coastal areas and peat as well (Astafiev *et al*. 2012).

The Bolshaya (Khibinian) Imandra depression is tectonically dominant and lies mainly within the Early Proterozoic Imandra-Varzuga Greenstone Belt. Ekostrovskaya Imandra and Babinskaya Imandra are located within the Archaean Belomorian Domain in a single area of Palaeozoic tectonic libialization. They show the west-eastern extension that conforms to the Tulioksky intratelluric strike-slip fault (5 in Fig. 2). Like Bolshaya Imandra, the Ekostrovskaya and Babinskaya Imandra depressions are tectonically prede-

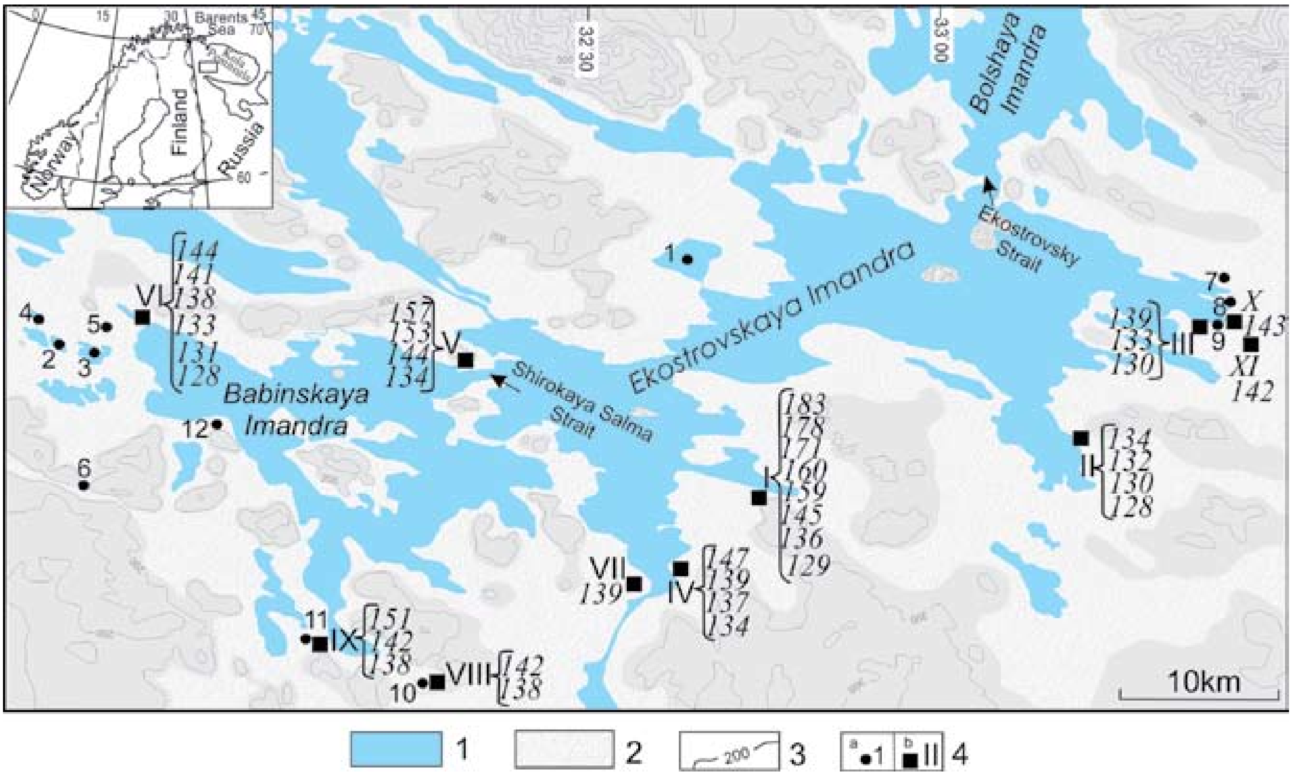

Fig. 1 Location map of the study area: 1– present-day Lake Imandra, 2 – coastal areas of Lake Imandra inundated with water in the Late Glacial and Holocene, 3 – elevation (100 m contour interval), 4 –studied localities with sedimentary sequences (a) and coastal terraces (b); terrace altitudes (m a.s.l.) modified after Armand *et al*. (1969) and Koshechkin *et al*. (1980) are shown in italics next to the curly bracket or black square

termined, but they were substantially rearranged by the Belomorian (White Sea) Ice Stream of the Fennoscandian Ice Sheet (FIS), which expanded there from west to south-east (Fig. 3) in the Late Valdaian (Late Weichselian). Most of the Bolshaya Imandra depression occurred in marginal areas of the Barentsevomorian (Barents Sea) Ice Stream and within the ice-divided zone of the Belomorian and Barentsevomorian Ice Streams of the FIS (Fig. 3). Based on the undated stacked record of subglacial ice flow traces, i.e. striations, eskers, till and bedrock lineations (drumlins), meltwater channels, changes in the FIS stream pattern caused by the shifting of the ice-divided zone to the south were identified (Hättestrand, Clark 2006; Winsborrow *et al*. 2010); the Barentsevomorian Ice Stream of the FIS existed in the Ekostrovskaya and Babinskaya Imandra areas during deglaciation prior to the Younger Dryas. During the final FIS degradation on the western Kola Peninsula, these areas are likely to have been deglaciated (Armand *et al*. 1964; Strelkov 1976; Yevzerov, Kolka 1993; Lenz et al. 2020).

MATERIALS AND METHODS

Coastal topography and bottom sediments from six small lake basins (Fig. 4) in coastal areas of southern Lake Imandra were studied for the palaeoenvironmental reconstructions of pra-Imandra in the Younger Dryas and early Holocene. Lithological, diatom and radiocarbon (^{14}C) dating data were obtained from the sedimentary sequences in small lake basins (localities 1–6 in Fig. 1). The bottom sediments cored in the deepest flat part of the lake bottom from the ice in winter and from the floating catamaran platform in summer. The cores were sampled with the overlapping of several centimetres, using a hand-operated piston corer that allowed obtaining material of undisturbed sedimentary sequence. The length and diameter of each core were 1.0 and 0.052 m, respectively. The sediments were described *in situ* according to their visually recognizable features (grain size of material, colour, structures). Samples of certain volume were taken with intervals specified by the appropriate method.

^{14}C dating was provided using the traditional scintillation method at the Laboratory for Geochronology and Geoecology of Bottom Sediments of St. Petersburg State University (LU samples) and at the Laboratory for Radiocarbon Dating and Electron Microscopy of the Institute of Geography RAS (IGAN sample) in line with the standard procedure (Arslanov 1987; Zazovskaya 2016). ^{14}C data were calibrated using the CalPal program (Danzeglocke *et al*. 2007). Table 1 presents the data obtained for the studied sedimentary sequences.

Diatoms were examined by standard methods (Proshkina-Lavrenko 1974; Gleser *et al*. 1992; Barinova *et al*. 2006) and classified as polyhalobous,

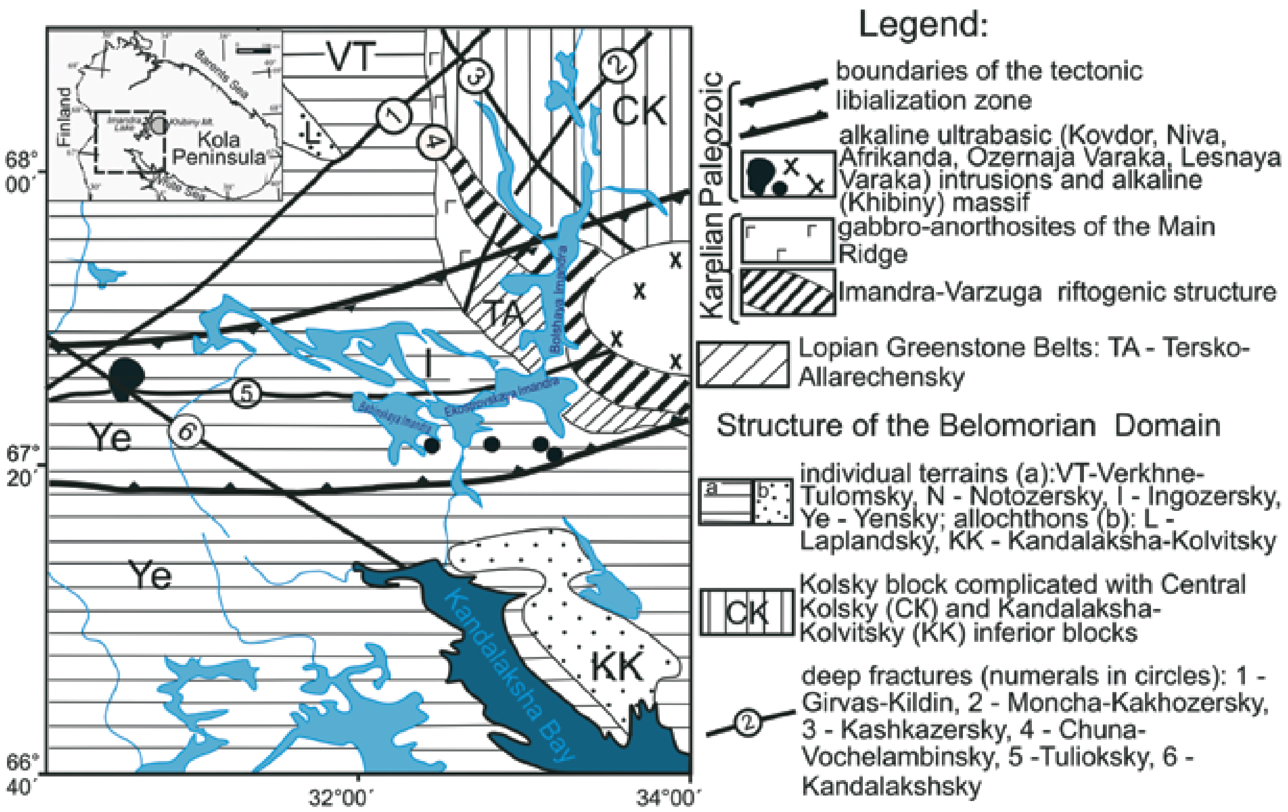

Fig. 2 Tectonic structure of the Lake Imandra depression and adjacent areas, modified after Astafiev *et al*. (2012)

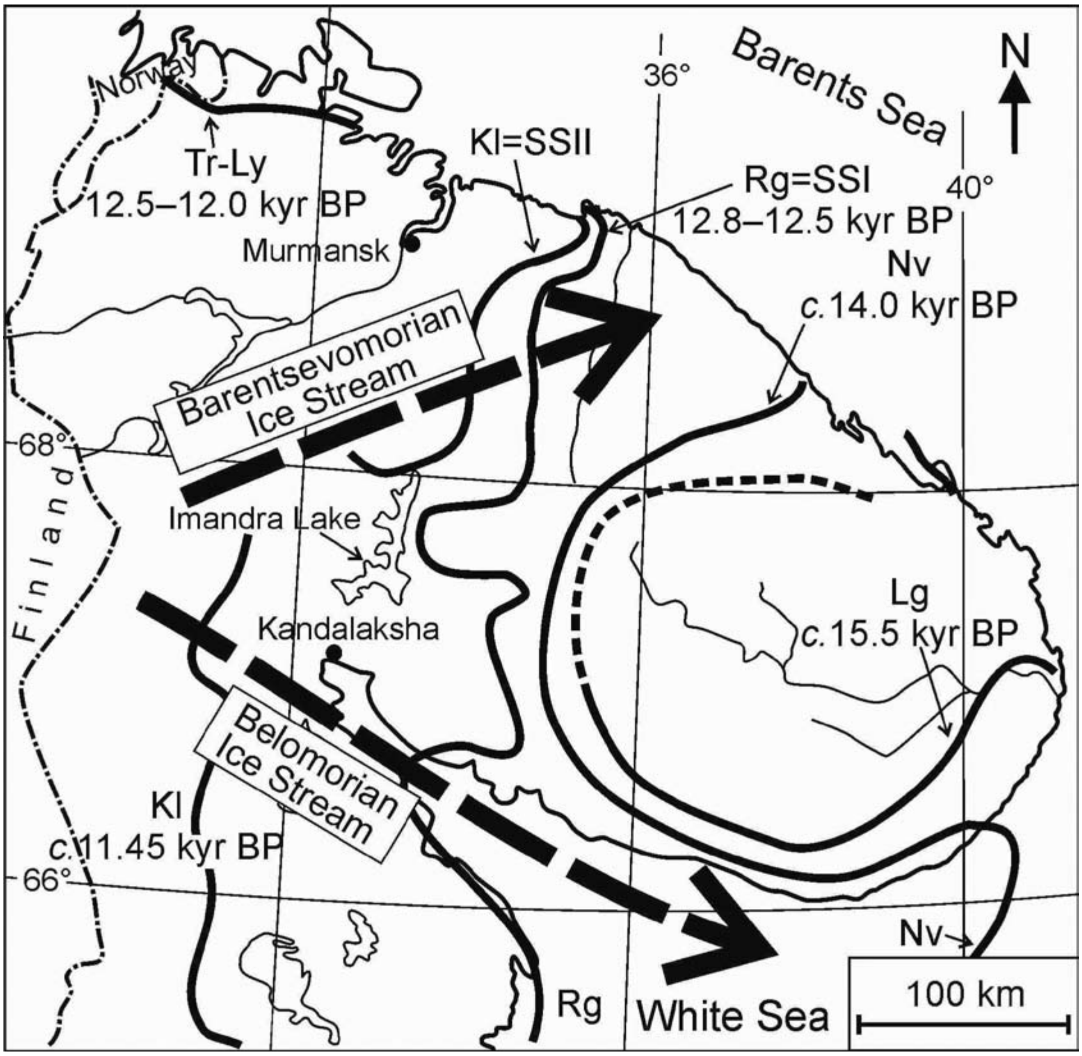

Fig. 3 Eastern flank of the Fennoscandian Ice Sheet during deglaciation. Ice marginal position marked by thick lines at different stages (Ekman, Iljin 1991): Lg – Luga (*c.* 15.5kyr BP, after Ekman, Iljin (1991)), Nv – Neva (*c.* 14.0 kyr BP, evidenced from dating of periglacial lake sediments in the south-eastern Kola Peninsula (Bakhmutov *et al.* 1994)), Rg and SSI – Rugozero and Salpausselkä I (early Younger Dryas, *c.*12.8–12.5 kyr BP, after Stroeven *et al.* (2016)), Tr-Ly – Tromsø-Lyngen (Younger Dryas, *c.* 12.5–12.0kyr BP, after Vorren, Plassen (2002)), Kl and SSII – Kalevala and Salpausselkä II (late Younger Dryas, *c.* 11.45kyr BP, after Putkinen et al. (2011)). Intermittent arrows show the main ice stream directions

Table 1 Radiocarbon dates obtained from bottom sedimentary sequences; lake basins are numbered according to Figs 1 and 4

Site No	Coordinates	Depth, m	Laboratory code	^{14}C date (yr BP)	Calibrated age (cal. yr BP) 2σ	Material dated	Reference
1	67°34′18″N 32°38′10″E	5.48–5.56	LU-6711	9750 ± 190	11156 ± 322	silty clay	Tolstobrova *et al.* 2016
		5.59–5.65	LU-6710	9820 ± 260	11317 ± 450	silty gyttja	
2	67°31′58.9″N 31°45′10.8″E	4.86–4.98	LU-7363	8300 ± 290	9210 ± 360	gyttja	Nikolaeva *et al.* 2016
		4.30–4.40	LU-7364	6490 ± 270	7330 ± 270	gyttja	
		3.95–4.05	LU-7365	5620 ± 300	6640 ± 340	wood	
3	67°31′37.3″N 31°48′00.0″E	6.26–.20	LU-7909	9090 ± 190	10250 ± 280	gyttja	Nikolaeva et al. 2016
4	67°32′59.6″N 31°42′53.8″E	6.26–6.19	LU-7575	9850 ± 120	11370 ± 190	gyttja	this paper
5	67°32′40″N 31°48′70″E	3.75–3.85	IGAN-4548	7700 ± 120	8491 ± 109	peat	Nikolaeva *et al.* 2015
6	67°27′60″N 31°46′08″E	4.10–4.25	LU-7368	8230 ± 300	9130 ± 360	gyttja	Nikolaeva *et al.* 2015

mesohalobous, oligohalobous halophiles oligohalobous indifferent and oligohalobous halophobes (Hustedt 1957). Diatom taxa names are provided in this paper according to a database of information on algae (Guiry, Guiry 2020).

The same results of palaeoseismological and palaeolimnological research (Nikolaeva *et al*. 2015; Tolstobrova *et al*. 2016) (localities 1–3 and 5, Figs 1 and 4) and same lithological and diatom data derived from earlier studies (Armand *et al*. 1964, 1969; Armand, Samsonova 1969) (localities 7–12, Figs 1 and 5) were also considered.

As identified on aerial photographs and large-scale maps, several coastal localities (indicated by Roman numerals in Fig. 1) provide coastal geomorphology data. Step-like landforms (terraces) were observed at

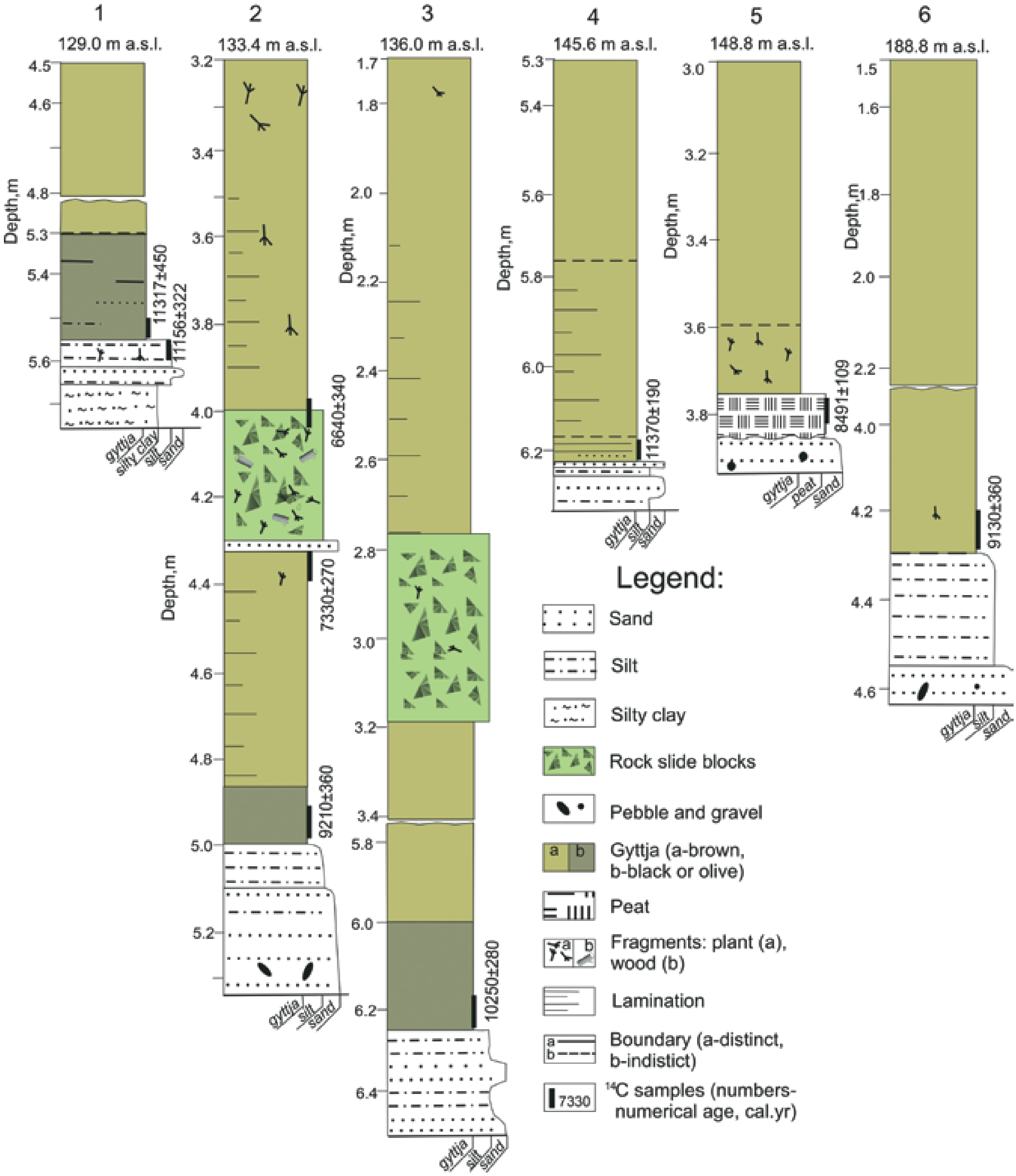

Fig. 4 Lithology logs of the bottom sediments cored in small lake basins in the Ekostrovskaya and Babinskaya Imandra coastal area

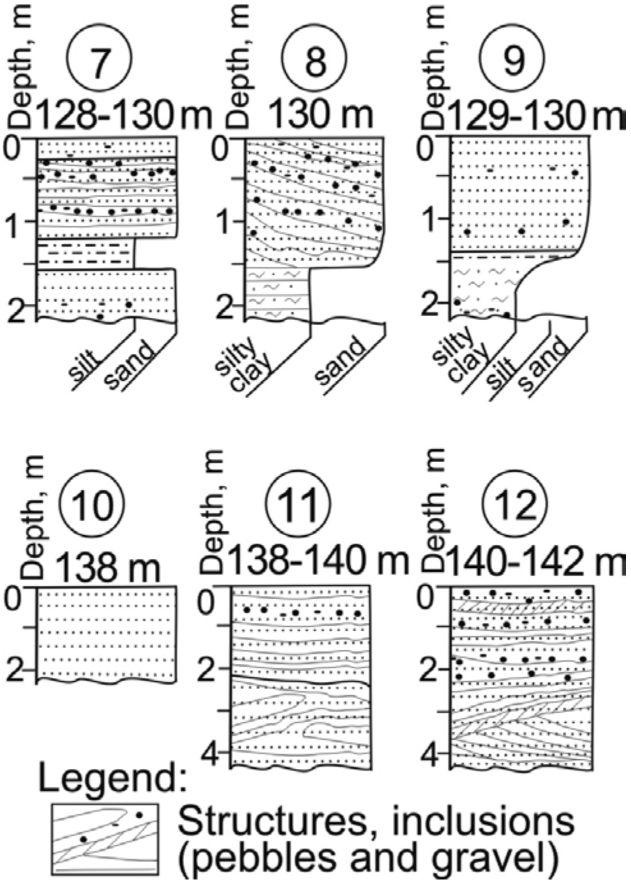

Fig. 5 Lithology logs of the sedimentary sequences from natural outcrops and sandpit on the Lake Imandra coast, modified after Armand *et al.* (1964, 1969) and Armand, Samsonova (1969)

different altitudes measured by levelling profiles (localities I–XI, Fig. 1).

The currently available evidence suggests certain palaeoenvironmental reconstructions in the coastal areas of Ekostrovskaya and Babinskaya Imandra.

COASTAL GEOMORPHOLOGY

Step-like landforms were identified in coastal areas north and south of Babinskaya Imandra, as well as south and east of Ekostrovskaya Imandra (Fig. 1). The youngest coastlines conform to lowermost terrace levels at the altitudes 128–129, 130–131 and 132 m a.s.l. in south-eastern and eastern coasts of Ekostrovskaya Imandra (Fig. 1, localities I and II, II and III, and II, respectively) and on the north-western coast of Babinskaya Imandra (locality VI, Fig. 1). Step-like landforms at the altitudes 133–134 m a.s.l. were also identified in localities II–VI (Fig. 1). On the south-western coast of Ekostrovskaya Imandra (localities I and IV, Fig. 1), a coastline at the altitudes 136–137 m a.s.l. was recognised by levelling profiles. Most ancient lacustrine coastlines are located at the altitudes 143–144 and 141–142 m a.s.l. in the south-western coastal areas and on the easternmost coast of Ekostrovskaya Imandra (localities VIII, and X–XI, Fig. 1), as well as on the northern and southern coasts of Babinskaya Imandra (V–VI and IX, Fig. 1).

A clearly exposed coastline was identified at 138–139 m a.s.l. on the eastern and south-western coasts of Ekostrovskaya Imandra (localities III, IV, VII, and VIII, Fig. 1) and in coastal areas of Babinskaya Imandra (localities VI and IX, Fig. 1). There are several step-like landforms from 145 to 183 m a.s.l. (localities I, IV, V and IX, Fig. 1), which seem to be of proglacial lake bench shape in the glaciofluvial sandy sediment. On the south-western coast of Ekostrovskaya Imandra (locality X, Fig. 1), the uppermost lacustrine coastline was identified at the altitudes 142 m a.s.l. (Armand, Samsonova 1969).

According to the topography of the highest lacustrine terraces in different coastal areas, the height of the lacustrine limit varies from the altitude of *c.* 143–145 m a.s.l. to *c.* 173–178 and 183 m a.s.l. (Fig. 1). Therefore, these coastline levels should be considered as limiting levels of the Lake Imandra coastal zone.

LITHOLOGY AND DIATOMS IN SEDIMENTARY SEQUENCES

Bottom sedimentary sequences from small lake basins at the altitudes of 129.0 to 188.0 m a.s.l. in the coastal area of Lake Imandra (localities 1–6, Fig. 1) consist of the lower minerogenic, or clastic, unit and the upper organogenic one (Fig. 4). Clastic units are composed of silty clay, silt and fine-grained sand in the lake basin 1, coarse-, medium- and fine-grained sand and silt in the lake basin 2, fine-grained sand and silt in the lake basin 3, medium-grained sand in the lake basin 4, fine-grained sand in the lake basin 5, as well as coarse-grained sand with pebbles and gravel, alternated silty clay and silt in the lake basin 6 (Figs 1 and 4). These sediments were deposited in a proglacial lake during deglaciation and in pra-Imandra Lake at the beginning of the Holocene.

Figure 4 shows that organogenic sequences consist of gyttja (sapropel) lying on organic-enriched silt. Lower parts of gyttja were ^{14}C dated between 11370 ± 190 and 9130 ± 360 cal. yr BP (Table 1). The sedimentary sequence from the lake basin 5 (148.8 m a.s.l.) hosts peat mixed with sand that was ^{14}C dated to 8490 ± 190 cal. yr BP (Fig. 4). In sequences from basins 2 (133.4 m a.s.l.) and 3 (136.0 m a.s.l.), the units with rock slide blocks were identified. These deformed layers consist of mixed sandy and silty fragments of different colour and size, plant debris and sandy patches in a sapropel matrix with the age of 6490 ± 270 cal. yr BP (Nikolaeva *et al.* 2016).

Figure 5 presents the lithology of sections in natural outcrops and sandpit on the eastern coast of Ekostrovskaya Imandra (localities 7–9, Fig. 1) and on the

Table 2 Diatoms in the sections from the southern coastal area of the Lake Imandra depression; localities are numbered according to Figs 1 and 5

Locality, m a.s.l. (number correlates to the locality number in Fig. 1)	Diatoms	References
7, c. 128–130: c. N 67.562061; E 33.262180, natural outcrop, abrasion slope of terrace; sand with silty interlayer (see Fig. 5)	Poly- and mesohalobous, *in situ*: *Isthmia nervosa* Kütz., *Paralia sulcata* (Ehr.) Cl., *Ellerbeckia clavigera* (Grun.) Crawford et Sims, *Hyalodiscus scoticu*s (Kütz.) Grun., *Grammatophora* sp., *Navicula distans* W. Sm. (Bréb.), *Campylodiscus echeneis* Ehr.ex Kütz., *Thalassiosira* sp., *Coscinodiscus* sp., *Podosira* sp. Poly- and mesohalobous, reworked (Cretaceous, Paleogene and Neogene): *Paralia sulcata* var. var., *Hyalodiscus kryshtofovichii* Jousé, *H.* sp. sp., *Stephanopyxis broschii* Grun., *St. grunowii* Gr. et St., *St.* sp. sp., *Coscinodiscus moelleri* A. Schmidt, *C. payeri* Grun., *Stellarima microtrias* (Ehr.) Hasle et Sims, *Hemiaulus polymorphus* Grun., *Trinacria* sp., *Isthmia* sp., *Pyxilla* sp., *Pseudopyxilla* sp., *Ktenodiscus aculeiferus* (Grun.) Blanco et Wetzel, *Goniothecium rogersii* Ehr., *Grunowiella gemmata* (Grun.) V.H., fragments of *Centrales*. Oligohalobous: *Paralia scabrosa* (Østr.) Moiseeva, *Aulacoseira islandica* subsp. *helvetica* (O.Müll.) Simons., *A. ambigua* (Grun.) Simons., *A. italica* (Ehr.) Simons., *Tetracyclus* sp. sp., *Staurosirella lapponica* (Grun.) Williams et Round, *Neidium iridis* (Ehr.) Cl., *Gyrosigma attenuatum* (Kütz.) Rabenh., *Eunotia* sp., *Stauroneis* sp., *Pinnularia borealis* Ehr., *P. viridis* Ehr., *Pinnularia* sp., *Epithemia* sp., *Cymbella* sp., *Didymosphenia geminata* (Lyngb.) M. Sch.	Armand *et al*. 1964
8, c. 130: c. N 67.535440; E 33.378909, natural outcrop, abrasion slope of esker; laminated silty clay (see Fig. 5)	Poly- and mesohalobous: *Paralia scabrosa* (Østr.) Moiseeva, fragments of *Centrales*; Oligohalobous: *Aulacoseira distans* (Ehr.) Simons., *A. italica* (Ehr.) Simons., *A. islandica* (O. Müll.) Simons., *A. islandica* subsp. *helvetica* (O.Müll.) Simons., *Eunotia* sp., *Ulnaria ulna* (Nitzsch) Compère, *Synedra* sp., *Pinnularia borealis* Ehr., *P. alpina* W. Sm., *P. lata* (Bréb.) W. Sm., *P.* sp. sp., *Epithemia* sp., *Rhopalodia gibba* (Ehr.) O.Müll., *Hantzschia amphioxys* (Ehr.) Grun., *H. amphioxys* var. *major* Grun.	Armand *et al*. 1964
9, c. 129–130: c. N 67.530008; E 33.335372, natural outcrop, abrasion slope of the hill; silty clay and silt (see Fig. 5)	Poly- and mesohalobous, *in situ*: *Paralia sulcata* (Ehr.) Cl., *Cocconeis costata* Greg. Poly- and mesohalobous, reworked (Paleogene and Neogene): *Hyalodiscus kryshtofovichii* Jousé, *Hyalodiscus* sp., *Diploneis bomboides* var. *moesta* (A.Schmidt) Cl. Oligohalobous, in situ: *Aulacoseira granulata* (Ehr.) Simons., *A. islandica* subsp. *helvetica* (O.Müll.) Simons., *Pinnularia borealis* Ehr., *Pinnularia* sp.	Armand *et al*. 1964
10, *c.* 138: *c.* N 67.337430; E 32.438758, natural outcrop, terrace in the nameless river valley; sand (see Fig. 5)	Poly- and mesohalobous, *in situ*: *Paralia sulcata* (Ehr.) Cl., *Diploneis interrupta* (Kütz.) Cl., *Diploneis* sp. Polyhalobous, reworked (Paleogene and Neogene): scarce fragments of *Centrales.* Oligohalobous: *Hantzschia amphioxys* (Ehr.) Grun., *Aulacoseira distans* (Ehr.) Simons., *Staurosirella martyi* (Hérib.-Joseph) Morales et Manoylov, *Eunotia faba* (Ehr.) Grun., *Pinnularia* sp., *Epithemia* sp.	Armand, Samsonova 1969
11, *c.* 138–140: *c.* N 67.380199; E 32.113573, natural outcrop, glaciofluvial delta on outwash plain; sand (see Fig. 5)	Poly- and mesohalobous*, in situ*: *Paralia sulcata* (Ehr.) Cl., *P. scabrosa* (Østr.) Moiseeva, *Coscinodiscus* sp., *Fragilariopsis oceanica* (Cl.) Hasle, fragments of *Centrales.* Oligohalobous, *in situ*: *Pantocsekiella kuetzingiana* (Thw.) Kiss et Ács, *Cyclotella meneghiniana* Kütz., *Stephanodiscus astraea* (Kütz.) Grun., *Aulacoseira distans* (Ehr.) Simons., *A. italica* var. *tenuissima* (Grun.) Simons., *Lindavia bodanica* (Eulen. ex Grun.) Nakov, Guill., Julius, Theriot et Alverson, *Cyclostephanos dubius* (Hust.) Round, *Tetracyclus glans* (Ehr.) Mills, *Tabellaria fenestrata* var. *geniculata* A. Cl., *T. flocculosa* (Roth.) Kütz., *Staurosirella martyi* (Hérib.) Morales et Manoylov, *S. pinnata* (Ehr.) Williams et Round, *Staurosira construens* Ehr., *Ulnaria ulna* (Nitzsch) Compère, *Eunotia bidentula* W. Sm., *E. faba* (Ehr.) Grun., *Skabitschewskia oestrupii* (A.Cl.) Kuliskovskiy et Lange-Bert., *Stauroneis* sp., *Cavinula pseudoscutiformis* (Hust.) Mann et Stickle, *Pinnularia* sp., *Reimeria sinuata* (W.Greg.) Kociolek et Stoermer.	Armand, Samsonova 1969
12, *c.* 140–142: *c.* N 67.490623; E 31.938068, side wall of sandpit, terrace; sand (see Fig. 5)	Poly- and mesohalobous, reworked: fragments of *Paralia sulcata* (Ehr.) Cl. vars., *P. scabrosa* (Østr.) Moiseeva and *Centrales.* Oligohalobous, *in situ*: *Cyclotella meneghiniana* Kütz., *Aulacoseira distans* (Ehr.) Simons., *A. granulata* (Ehr.) Simons., *A. italica* (Ehr.) Simons., *Cyclostephanos dubius* (Hust.) Round, *Tetracyclus glans* (Ehr.) Mills, *Tabellaria flocculosa* (Roth.) Kütz., *Staurosirella martyi* (Hérib.) Morales et Manoylov, *S. pinnata* (Ehr.) Williams et Round, *Staurosira construens* Ehr., *Ulnaria ulna* (Nitzsch) Compère, *Eunotia* sp., *Pinnularia borealis* Ehr., *P. viridis* (Nitzsch) Ehr.	Armand *et al*. 1969

south-western coast of Babinskaya Imandra (localities 10–12, Fig. 1). The sections are located in an outcrop on abrasion slope of the coastal terrace (locality 7, Fig. 5), esker (locality 8, Fig. 5) and moraine hill underlain by marine sediments (locality 9, Fig. 5). These sedimentary sequences are made of marine and glaciolimnic sand with a silty interlayer, glaciolimnic laminated silty clay covered with glaciofluvial sand, and marine silty clay overlapped with glacial sand and silt, respectively. Marine and lacustrine sandy sediments were identified in the outcropped slopes of river and lake terraces (localities 10 and 12, Figs 1 and 5) and in the glaciofluvial delta (locality 11, Figs 1 and 5). The marine and lacustrine genesis of sediments was evidenced by the diatom data (Armand et al. 1964, 1969; Armand, Samsonova 1969) summarized in Table 2.

Diatoms were studied by D.B. Denisov and A.L. Kosova (Nikolaeva *et al*. 2015, 2016) in the organogenic unit from lake basins 2 (133.4 m a.s.l.) and 5 (148.8 m a.s.l.) in the westernmost coastal area of the Lake Imandra depression (Figs 1 and 4). No poly- or mesohalobous diatoms were identified in the sediment from these sections. The diatom assemblages in the lowermost gyttja sequence from the basin 5 (148.8 m a.s.l.) mainly host benthic and periphytic oligohalobous species, planktonic diatoms are rare. *Brachysira* sp., *Frustulia saxonica* Rabenh., *Planothidium minutissimum* (Krasske) E.A. Morales, *P. lanceolatum* (Breb. ex Kutz.) Lange-Bertalot, *Staurosirella pinnata* (Ehrb.) Williams et Round, *S. pinnata* var. *intercedens* (Grun.) P.B. Hamilton, *Staurosira construens* Ehrb. are the most common species in records. In mixed sapropelic, sandy and silty sediments of the deformed layer from the basin 2 (133.4 m a.s.l.), the total diatom abundance abruptly decreases three times at the depth of 4.0–4.3 m (Fig. 4), preserving the same species diversity. On the deformed layer, the 25 cm-thick gyttja contains planktonic *Aulacoseira alpigena* (Grun.) Kramm. and *A. distans* (Ehrb.) Simons. that show the peak abundance and become dominant taxa there. It results from a dramatic change in certain habitats, which appears to be caused by an earthquake and disturbance in sediment accumulation.

Scarce poly- and mesohalobous and oligohalobous halophiles were identified in the lower part of sedimentary sequences in the outcrops on the coastal Ekostrovskaya Imandra and southern Babinskaya Imandra (localities 7–12, Figs 1, 4 and 5) and in the cores from the Lake Osinovoe basin at 129 m a.s.l. (locality 1, Figs. 1 and 4). In the sediment record from Lake Osinovoe, a total of 145 diatom taxa were identified. The lowermost clayey sediments (Zone Ia in Fig. 6) are barren of diatoms. Mesohalobous *Diploneis pseudovalis* Hust. and halophilic *Staurosirella pinnata* (Ehrb.) Williams et Round were recognized in the superposed sandy deposits (Zone Ib in Fig. 6), suggesting their sedimentation in a brackish-water reservoir. Diatoms demonstrate a rise in abundance in the lowermost organic sequence (Zone II in Fig. 6), which is characterized by a higher percentage of oligohalobous indifferent, such as *Fragilaria construens* (Ehrb.) Hust., *Pseudostaurosira brevistriata* (Grun.) Williams et Round, *Staurosira venter* (Ehrb.) Kobayasi. Halophilic *Staurosirella pinnata* (Ehrb.) Williams et Round shows peak values (82%) there. These pioneer diatom species have the ability to adapt to rapidly changing environment and tolerate poor lighting conditions. They are commonly attributed to the Late Glacial fresh-water sediments (Smol 1988; Wolfe 1996; Grönlund, Kauppila 2002). The diatom proxy, lithostratigraphy and ^{14}C data (Table 1, Figs 4 and 6) suggest that the brackish-water reservoir turned into fresh-water at the end of the Younger Dryas – beginning of the Preboreal, no later than *c.* 11,400 cal. yr BP. A pronounced rise in diatom abundance is recorded in the superposed gyttja (Zone III in Fig. 6), particularly in the occurrence of periphitic taxa, such as dominant *Fragilariforma, Staurosira, Staurosirella* and *Pseudostaurosira*; the value of planktonic *Cyclotella* reaches 11% there. Upwards in the gyttja layer (Zone IV in Fig. 6), not only high abundance of *Cyclotella* was recorded in the species composition, but also other planktonic taxa were identified, such as *Aulacoseira*. Meanwhile, benthic *Brachysira, Caloneis, Diploneis, Navicula, Neidium, Pinnularia, Eucocconeis, Achnanthidium, Psammothidium, Cymbella, Eunotia, Gomphonema* and dominated *Brachysira zellensis* (Grun.) Round et Mann (7%) are highly abundant within Zone IV. Co-existence of benthic and planktonic diatom taxa indicates relatively deep and clear water conditions in the reservoir. In addition, the local diatom complex occur restructures with respect to the pH, which is reflected in increase in the share of indifferent and acidophilic species. This suggests isolation of the Lake Osinovoe basin from Lake Imandra. In a small water reservoir isolated from Lake Imandra, a slow decrease in pH could be caused by natural acidification resulting from humic acid input following catchment paludification. The uppermost gyttja related to the late Holocene (Zone V in Fig. 6) hosts oligohalobous indifferent and halophobous, bentic, periphitic and planctonic diatoms, such as dominant *Nupelavitiosa* (Schimanski) Siver et Hamilton, sub-dominated *Aulacoseira valida*, *Cyclotella radiosa*, *C. rossi* Håkans., *Eunotia diodon* Ehrb., *Sellaphora laevissima* (Kütz.) Mann and *Staurosira venter*.

DISCUSSION

The presented lithological, ^{14}C and diatom data from sediment sequences provide an interpretive framework of palaeoenvironmental history of the

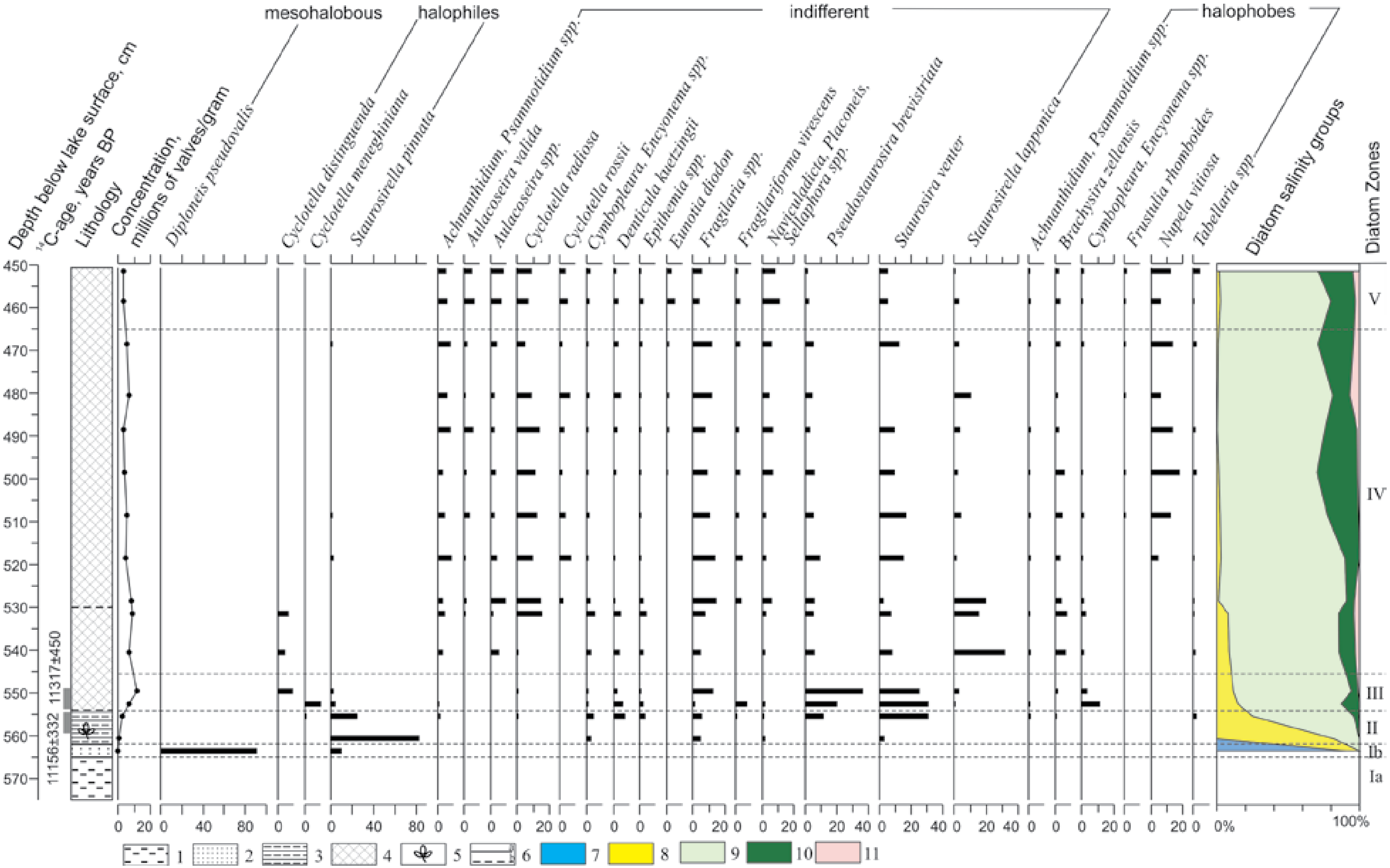

Fig. 6 Percentage diagram of the selected diatom taxa relative to the diatom sum and diatom salinity groups, from the sediment record of Lake Osinovoe (1 in Figs 1 and 4), analyst Tolstobrova A. 1 – silty clay, 2 – sand, 3 – silty gyttja, 4 – gyttja, 5 – plant macroremains, 6 – contacts (a – abrupt, b – gradual); diatoms salinity groups: 8 – mesohalobous, 9 – oligohalobous halophiles, 10 – oligohalobous indifferent, 11 – oligohalobous halophobes, 12 – unknown

southern Lake Imandra depression in the Younger Dryas and early Holocene. Including the territory, which was the bottom floor of the periglacial basin followed by pra-Imandra Lake, the coastal area of Ekostrovskaya and Babinskaya Imandra is confined to the highest terraces at the altitudes from *c.* 143–145 m a.s.l. to *c.* 173–178 and locally 183 m a.s.l. The different contemporaneous location appears to be caused by the Late Glacial and Postglacial tectonic activity in the fracture zones (Nikolaeva *et al.* 2018), i.e. Tulioksky fault (5 in Fig. 2), and different tectonics uplift of the local geological structures, i.e. Palaeozoic intrusions in the locality I (Figs 1 and 2). In addition, since the study area occurs in the marginal zone of the Late Valdaian (Late Weichselian) ice sheet (Fig. 3), the measured altitudes of the coastline fragments (localities I–XI, Fig. 1) also seem to indicate some differences in glacioisostatic rebound.

Tectonic or, specifically, seismotectonic activity along the Tulioksky deep fracture (5 in Fig. 3) and multidirectional coseismic movements are assumed to have induced a catastrophic sediment slumping from the basin sides and the occurrence of deformed layers in sedimentary sequences from the lake basins 2 (133.4 m a.s.l.) and 3 (136.0 m a.s.l.) (Figs 1 and 4). In that event, some of the sediments deposited on the basin sides could suddenly slip down into the deeper part of the reservoir. Seismogenic shaking and "cracking" of sediments caused suspension of debris, which was further redeposited onto an undeformed layer. Similar deformations in sedimentary sequences are often produced by lacustrine seiches or tsunamis and associated with earthquakes (Dawson, Stewart 2007; Heifetz *et al.* 2005). ^{14}C data 7330 ± 270 and 6440 ± 340 cal. yr BP from the studied sediment sequence (lake basin 2, Figs 1 and 4, Table 1) suggest the middle Holocene age of the catastrophic event in the Lake Imandra basin.

Lithostratigraphical and diatom data suggest that the Late Glacial – early Holocene environment varies in the Ekostrovskaya and Babinskaya Imandra depressions, reflecting a different deglaciation pattern and differences associated with marine water penetration and fresh-water inflow into that reservoir. As summarized by Stroeven *et al.* (2016), all available deglaciation patterns for the Fennoscandian Ice Sheet imply that the central Kola region, including the Lake Imandra depression (Fig. 3), was glaciated in the Younger Dryas (12.7–11.5 cal. kyr BP). Our lithostratigraphical, ^{14}C and diatom data on the Lake Osinovoe (1, Fig. 1) sediment sequence (Fig. 6, Table 1) suggest that the brackish-water reservoir followed by the fresh-water one existed in the Ekostrovskaya Imandra Lake depression earlier than 11.3 cal. kyr BP, i.e. in the second half of the Younger Dryas. In addition, fresh-brackish-water, brackish-water and scarce marine diatoms (Table 2, and Fig. 6) were identified in the lower part of sedimentary sequences in the outcrops on the eastern Ekostrovskaya Imandra and southern Babinskaya Imandra coastal areas (localities 7–11, Figs 1 and 5). In the sequences studied, marine or brackish-water sand, silt and silty clay with oligohalobous halophiles, meso- and polyhalobous diatoms occur up to 136–138 m a.s.l. (Armand *et al.* 1964, 1969; Armand, Samsonova 1969). Only oligohalobous diatoms were found in the sequences from the western Babinskaya Imandra coastal area (locality 2, 5, and 12, Figs 1, 4 and 5; Table 2). No diatom proxies of marine ingression into western Lake Imandra depression were recorded there.

There is an opinion (Kolka *et al.* 2005; and others) that an extensive periglacial freshwater lake existed in the Kandalakshsky Bay and adjacent areas included the Imandra Lake depression in the Allerød *c.* 14.0–13.0 cal. kyr BP (Fig. 7a). The results derived from bottom sediments cored in coastal lakes on the Kandalaksha Bay area (Lunkka *et al.* 2012) show that an extensive ice lake existed in the White Sea basin prior to 12.0–11.8 cal. kyr BP. During the Late Glacial marine transgression, saline water started inundating into the White Sea basin at the end of the Allerød. Lithostratigraphical and diatom results, together with dating and spore-pollen proxies, indicate that marine conditions occurred in the Kandalaksha Bay in the early part of the Younger Dryas *c.* 12.3 kyr BP (Korsakova *et al.* 2016). Vast areas to the west of the present-day White Sea basin were submerged with periglacial lakes during and after the Younger Dryas Stadial *c.* 11.5–11.3 kyr BP (Pasanen *et al.* 2010). The current results imply that the FIS ice margin was located in the Lake Imandra depression during the Younger Dryas Stadial earlier than *c.* 11.4 cal. kyr BP; thus the ice sheet covered the western Babinskaya Imandra, and periglacial brackish-water reservoir existed in the Ekostrovskaya and eastern Babinskaya Imandra depressions and in the White Sea depression *c.* 12.3 kyr BP (Fig. 7b). The Younger Dryas cooling and the decreased meltwater input were favourable for increase in water salinity in that periglacial basin. We suggest that periglacial pre-Imandra Lake reservoir was connected with the White Sea marine reservoir for a short period of time. Given the extremely low ice surface gradients and the following warm climate, a rapid ice-marginal retreat after its Younger Dryas expansion could occur continuously, with no prolonged stillstand in the Lake Imandra depression and adjacent areas. A discernible deglaciation record is substantially presented there from meltwater landforms (Hättestrand, Clark 2006; Stroeven *et al.* 2016). During the Younger Dryas Stadial, the eastern Lake Imandra depression seemed to be deglaciated; silty clay, silt and sand composed the lower minerogenic units in the bottom sedimentary sequences (1–6, Figs 1 and 4) deposited under glaciofluvial and glaciolacustrine conditions. In the coastal areas, these sediments occur in the glaciofluvial deltas (locality 11, Figs 1 and 5), lake terraces (localities 10 and 12, Figs 1 and 5), eskers (locality 8, Figs 1 and 5), and other meltwater landforms. During deglaciation, intraglacial reservoirs, succeeded by fresh-water lakes in the early Holocene *c.* 11.4–8.5 kyr BP locally occurred outside the western Lake Imandra coastal areas (lake basin 6, Figs 1 and 4) or nearest to the lake coastline (lake basins 4 and 5, Figs 1 and 4).

In the early Holocene *c.* 11.4–8.5 kyr BP, the pra-Imandra Lake coastline was at least 16–18 m higher than the modern one (Fig. 7c), as suggested by terrace altitudes (Fig. 1) and the position of the peat unit in the sedimentological sequence from the lake basin 5 (Figs 1 and 4) at 143–145 m a.s.l. Due to the lake level regression in the early Holocene, the basins 1–3 at the altitudes of up to 136.0 m a.s.l. (Figs 1 and 4), which used to be lagoons before (c, Fig. 7), were successively isolated from pra-Imandra Lake at *c.*10,200–9,200yr BP.

CONCLUSIONS

1. The coastal area of the southern Lake Imandra depression is confined to the highest terraces at the altitudes of *c.* 143–145 m a.s.l. to *c.* 173–178 and locally to 183 m a.s.l. The varied contemporary position of a high lacustrine limit is supposed to be caused by the Late Glacial and Postglacial tectonic activity in fracture zones and different tectonic uplifts of the local geological structures. Besides, the measured coastline-fragment altitudes seem to indicate a glacioisostatic rebound.

2. The eastern Lake Imandra depression seems to have been deglaciated in the Allerød–Younger Dryas. The FIS ice margin is assumed to have been in the Lake Imandra depression during the Younger Dryas Stadial, when the ice sheet covered western Babinskaya Imandra. Our lithostratigraphical, ^{14}C and diatom data suggest that the brackish-water reservoir followed by the fresh-water one existed in the Ekostrovskaya Imandra depression earlier than *c.* 11.4 cal. kyr BP, i.e. in the Younger Dryas.

3. In the Preboreal, the ice margin significantly retreated westward in the western Lake Imandra depression and adjacent areas. The extensive fresh-water pra-Imandra Lake basin formed there at the end of the Preboreal. In the Holocene, earlier than *c.* 8.5 cal. kyr BP, the pra-Imandra Lake coastline was at least 16–18 m higher than the modern one, as can be assumed according to coastal morphology and lithostratigraphic data. During the early Holocene, the coastline changed, water square reduced, and isolated small lakes occurred on the Lake Imandra coasts.

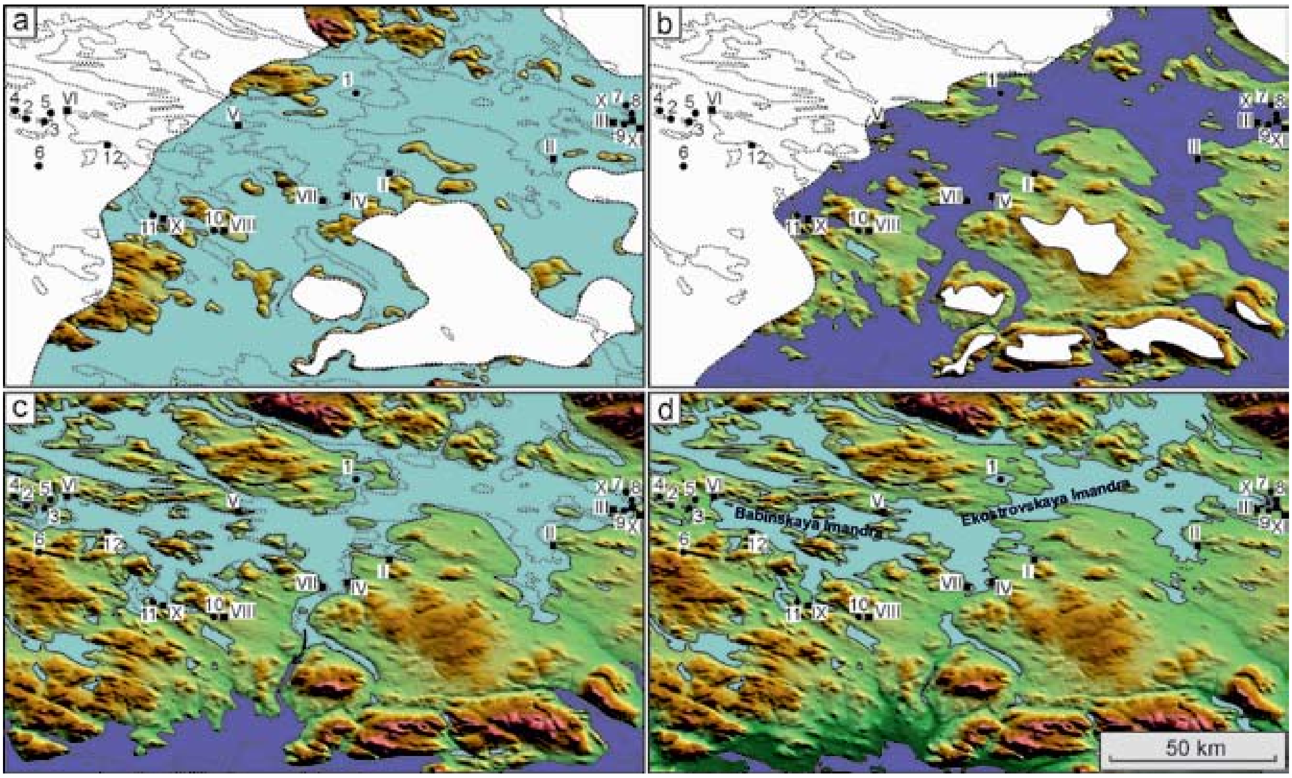

Fig. 7 Late Glacial and Postglacial setting in the Lake Imandra depression: (**a**) stage of the fresh-water proglacial lake (marked in blue), c. 14.0–13.0 kyr BP; (**b**) stage of the brackish-water periglacial basin (marked in dark-blue), *c.* 12.3 kyr BP; (**c**) stage of the fresh-water pra-Imandra Lake isolated from brackish-water basin, *c.* 11.4–8.5 kyr BP; (**d**) current setting in the Lake Imandra depression. The glacier is shown in white; dotted lines indicate the modern coastline of lakes and the White Sea; black circles and squares with numbers correspond to those in Fig. 1

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