

THESIS ON NATURAL AND EXACT SCIENCES B198

**Diatom Stratigraphy and Relative
Sea Level Changes of the Eastern Baltic Sea
over the Holocene**

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This dissertation was accepted for the defence of the degree of Doctor of Philosophy in Earth Sciences on November 19, 2015.

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Declaration: *Hereby I declare that this doctoral thesis, my original investigation and achievement, submitted for the doctoral degree at Tallinn University of Technology has not been submitted for doctoral or equivalent academic degree.*

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ISSN 1406-4723

ISBN 978-9949-23-872-9 (publication)

ISBN 978-9949-23-873-6 (PDF)

LOODUS- JA TÄPPISTEADUSED B198

Ränivetikate stratigraafia ja Läänemere idaosa veetaseme muutused Holotseenis

IEVA GRUDZINSKA

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LIST OF ORIGINAL PUBLICATIONS

I Grudzinska, I., Saarse, L., Vassiljev, J., Heinsalu, A., Veski, S. 2012. A palaeocoastline reconstruction for the Käsnu and Pärисpea peninsulas (northern Estonia) over the last 4000 years. *Estonian Journal of Earth Sciences* 61: 307–316. doi: 10.3176/earth.2012.4.09

II Grudzinska, I., Saarse, L., Vassiljev, J., Heinsalu A. 2013. Mid- and late-Holocene shoreline changes along the southern coast of the Gulf of Finland. *Bulletin of the Geological Society of Finland* 85: 19–34.

III Grudzinska, I., Saarse, L., Vassiljev, J., Heinsalu, A. 2014. Biostratigraphy, shoreline changes and origin of the Limnea Sea lagoons in northern Estonia: the case study of Lake Harku. *Baltica* 27: 15–24. doi: 10.5200/baltica.2014.27.02

IV Vassiljev, J., Saarse, L., **Grudzinska, I.**, Heinsalu, A. 2015. Relative sea-level changes and development of the Hiiumaa Island, Estonia, during the Holocene. *Geological Quarterly* 59: 517–530. doi: 10.7306/gq.1227

V Grudzinska, I., Vassiljev, J., Saarse, L., Veski S. (submitted). Diatom and sedimentary record of past environmental change and paleo sea water intrusions into coastal Lake Lilaste, Latvia. *Journal of Paleolimnology*.

ABBREVIATIONS

a.s.l. – above the present sea level

AL – Ancylus Lake

AMS – Accelerator Mass Spectrometry

AR – accumulation rate

b.s.l. – below the present sea level

BIL – Baltic Ice Lake

BP – ¹⁴C years before present (AD 1950)

cal BP – calibrated years before present (AD 1950)

DAR – diatom accumulation rate

DTM – digital terrain model

ESL – eustatic sea level

GIS – geographic information system

LIDAR – Light Detection and Ranging

LimS – Limnea Sea

LitS – Litorina Sea

LOI – loss-on-ignition

MM – mineral matter

MS – magnetic susceptibility

OM – organic matter

RSL – relative sea level

YS – Yoldia Sea

1. INTRODUCTION

In the 19th century, the cornerstone principle of geology was the uniformitarian concept that natural processes occurring today have functioned throughout most of the Earth's history. Charles Lyell in his seminal book *Principles of Geology* (Lyell 1853), based on uniformitarian ideas, stated that 'the present is the key to the past'. In 20th century, however, global climate change challenged this concept. Scientists in the 1980s determined that reversing the direction of the previously held theory enabled forecasting of future changes based on palaeoenvironmental data (Issar 2010).

Following the concept of 'the past is the key to the future', geologists have examined the development of the Baltic Sea. The Baltic Sea is a semi-enclosed brackish-water basin considered to be relatively vulnerable to the climate change and anthropogenic impacts. Therefore, research of palaeoshoreline displacement and factors of relative sea level (RSL) fluctuations in the past is a prerequisite for discussing future global climate change and predicting its processes in the coastal zones of the Baltic Sea. Moreover, palaeoreconstruction provides additional information for stakeholders in planning sustainable development of the coastal zone, and the results are useful for archaeologists in the investigation of ancient dwelling sites that were once situated along the palaeoshoreline (Jussila & Kriiska 2004; Miller et al. 2004; Veski et al. 2005; Jöns 2011; Rosentau et al. 2013).

From a geological point of view, interaction of the glacial and inter-glacial epochs over the Pleistocene and the Holocene has caused the Baltic Sea basin to undergo complicated transformations. Its history since deglaciation can be divided into the following five major Baltic Sea stages: the freshwater Baltic Ice Lake (BIL), the slightly brackish Yoldia Sea (YS), the freshwater Ancylus Lake (AL), the marine/brackish Litorina Sea (LitS), and the brackish Limnea Sea (LimS). The transitions from one stage to the next were provided and influenced by interplay between glacial isostatic land uplift and eustatic global sea level changes. Currently, the northern part of the Baltic Sea is still experiencing land uplift whereas the southern part is subject to submergence; the area around the Gulf of Bothnia is rising 8–9 mm yr⁻¹, and that around the southernmost Baltic Sea is sinking approximately 1 mm yr⁻¹ (Ekman 1996; Björck et al. 2008). The present-day apparent uplift rate in the eastern part of the Baltic Sea is relatively slow. In the coastal area of Estonia it varies from 2.5 mm yr⁻¹ on the Island of Hiiumaa to 0.5 mm yr⁻¹ in the surrounding region of Narva (Torim 2004). In the southern coast of the Gulf of Riga, however, the apparent land uplift rate is about 0 mm yr⁻¹ (Ekman 1996).

The shore displacement of the various Baltic Sea stages in the eastern Baltic has been studied by many authors. Ramsay (1929), considered to be the pioneer of such research, described the AL and the LitS beach formations in Estonia and reconstructed their isobases with values valid to the present. Numerous studies have followed (Kents 1939; Kessel & Raukas 1967, 1979; Ratas 1976; Raukas & Ratas 1996; Raukas et al. 1996; Königsson et al. 1998; Hang & Kokovkin 1999;

Saarse et al. 2003). Rosentau et al. (2009), Vassiljev et al. (2011), and Vassiljev & Saarse (2013) studied the development of BIL; Heinsalu (2000) and Heinsalu & Veski (2007) studied the YS; Saarse et al. (1999) studied the YS and AL; and Saarse et al. (2000, 2009a) and Rosentau et al. (2011, 2012) studied LitS. Despite several studies on the shoreline displacement in northern and western Estonia, no profound relative water-level curve of LimS has been reported on the basis of multi-proxy data.

Although the history of the Baltic Sea in Latvia has been studied by several authors (Ulsts 1957; Grinbergs 1957; Saule-Sleinis 1960; Aboltina-Presnikova 1960; Veinbergs 1979, 1996; Veinbergs et al. 1980; Eberhards 2000, 2006), their conclusions on the spatial and temporal distribution of the Baltic Sea shoreline were based mainly on morphological, lithological, and pollen data with limited numbers of radiocarbon dates. Saarse et al. (2000) stated that lack of biostratigraphic proxies indicating water environment, such as diatom analysis, and reliable radiocarbon ages often leads to misinterpretation of ancient coastal formations and their chronology. Therefore, eight lakes from the surroundings of Riga were studied in order to revise the RSL changes at the southern coast of the Gulf of Riga based on new litho- and biostratigraphical data and radiocarbon datings.

One of the most controversial questions among scientists in the field of the Baltic Sea development is the number of LitS transgressions. In southern Finland and northwestern Russia, one main LitS transgression has been described (Eronen et al. 2001; Miettinen 2002, 2004; Miettinen et al. 2007), whereas several waves of LitS transgression have been recognised in southeastern Sweden (Yu 2003; Berglund et al. 2005, Wohlfarth et al. 2008). According to early studies (Kents 1939; Thomson 1939; Kessel 1960; Kessel & Raukas 1979), multiple LitS transgressions in Estonia have been detected. However, the latest studies describe only one uniform transgression (Raukas et al. 1995; Veski et al. 2005; Saarse et al. 2009b). Yet Veinbergs (1996) and Eberhards (2000, 2006) discussed the evidence confirming twofold LitS transgression in the area of Latvia. Thus, no conclusive evidence of LitS transgression in the eastern Baltic area has been provided thus far.

The coastal area of the eastern Baltic Sea is rather flat; therefore, even minor sea-level fluctuations can be traced in sediment sequences of isolation basins. Moreover, ancient coastal formations of the various Baltic Sea stages are easily distinguishable in relief maps in northern and northwestern Estonia and the Island of Hiiumaa, particularly those produced by Light Detection and Ranging (LIDAR). However, at the southern coast of the Gulf of Riga, in which the uplift rate is close to zero, various environmental processes throughout time have eroded and relocated the sediments and have changed the coastal landforms of the previous Baltic Sea stages.

Diverse methods of sea-level reconstructions and limitations of sea-level archives have been discussed by Gehrels & Shennan (2015). The authors consider that accurate sea-level index points can be established by dating the stratigraphic contact at which the marine diatom assemblage is replaced by freshwater diatoms

or vice versa and by measuring the altitude of the threshold, which is the lowest point in the topography that separates the basin from the sea. These isolated coastal water bodies provide an excellent sedimentary archive of the evolutionary stages of this coastal region.

In summary, this thesis addresses three problematic questions concerning the development of the Baltic Sea in the eastern Baltic region: the poorly constrained shoreline changes of LimS in Estonia, the incomplete understanding of shoreline displacement of LitS on the Island of Hiiumaa, and the outdated knowledge of shoreline displacement and water-level fluctuations in Latvia, which are based on individual studies with insufficient numbers of radiocarbon datings and biostratigraphical data.

To resolve these issues, isolation basins in northern and northwestern Estonia, at the Island of Hiiumaa, and at the southern coast of the Gulf of Riga were studied by using the isolation method, which is widely applied in water-level reconstructions (Gehrels & Shennan 2015). This is the first attempt for litho-, bio- and chronostratigraphic study of isolation basins in the surrounding areas of Riga and for reconstructing the water-level curve of the southern coast of the Gulf of Riga.

The hypotheses of the thesis include the following points:

- a) High-resolution studies of the isolation basin sediment record provide good data for reconstruction of water-level curves to enable the detection of the relative sea-level fluctuations and patterns of the land uplift in areas both with higher land uplift rates and those with close to zero uplift.
- b) In the areas with high land uplift, such as the northern part of the Baltic Sea, evidence exists of only one prominent and relatively short LitS transgression with no significant water-level fluctuations afterwards. In the areas of slow uplift or near-zero isobase, however, such as the southern part of the Gulf of Riga, it is possible to observe time transgressive LitS highstand with minor water-level oscillations.
- c) The maximum water level during the YS and the AL stages in the areas with land uplift near 0 mm yr^{-1} was below the present sea level, and the LitS transgression was close to the present sea level.

The aims of the present thesis include the following objectives:

- 1) to investigate the palaeoenvironment of the isolation basins situated at various elevations with a focus on their isolation events and to reconstruct reliable water-level curves of LitS and LimS in northern and northwestern Estonia and at the Island of Hiiumaa;
- 2) to obtain new high-resolution data from isolation basins in the surrounding areas of Riga in order to determine the index points for reconstruction of the Holocene RSL curve in the southern part of the Gulf of Riga and to discuss the efficiency of diatom analysis for describing the magnitude of LitS transgression in areas with near-zero land uplift;

- 3) to analyse the similarities and differences in water-level fluctuation during the Holocene in areas with positive land uplift and those with near-zero isobase;
- 4) to produce relevant data for the Holocene Baltic Sea shoreline database (Saarse et al. 2003) for application to palaeoshoreline reconstruction; and
- 5) to discuss palaeogeography, shore displacement, and isostatic uplift patterns in the eastern Baltic Sea and adjacent regions.

The outcomes of this thesis are approbated in four publications in ISI Web of Science journals such as *Estonian Journal of Earth Sciences*, *Bulletin of the Geological Society of Finland*, *Baltica*, and *Geological Quarterly* and in one article submitted to *Journal of Paleolimnology*. The results were introduced at the following international conferences: the 22nd Nordic Diatomists' Meeting, Ratnieki, Latvia (20–22 May 2015), the 12th Colloquium on Baltic Sea Marine Geology, Rostock-Warnemünde, Germany (8–12 September 2014), the 21st Nordic Diatomists' Meeting, Norrtälje, Sweden (20–22 November 2013), Submerged Prehistoric Archaeology and Landscapes of the Continental Shelf (SPLASHCOS)–Under the Sea: Archaeology and Palaeolandscapes Final Conference, Szczecin, Poland (23–27 September 2013); the 2nd Conference of Doctoral School of Earth Sciences and Ecology 'Down to Earth', Tallinn, Estonia (16–17 May 2013), the 11th Colloquium on Baltic Sea Marine Geology, Helsinki, Finland (19–21 September 2012), the 12th International Paleolimnology Symposium, Glasgow, United Kingdom (21–24 August 2012), and the 20th Nordic Diatomists' Meeting at the Geological Survey of Denmark and Greenland (GEUS), Copenhagen, Denmark (23–25 May 2012).

2. THE DEVELOPMENT OF THE BALTIC SEA

The Baltic Sea is one of the largest brackish-water basins on the Earth. It developed into its present shape after the Late Weichselian glaciation and hence is rather young from geological point of view. As stated in the previous chapter, the combination of glacio–isostatic land uplift and eustatic sea-level rise has caused the Baltic Sea to undergo five complex postglacial development stages: BIL, YS, AL, LitS, and LimS. It should be noted that the Mya Sea is an additional stage mentioned in the literature (Munthe 1910; Hessland 1945). This stage covers the last 500 years since the soft sediment bivalve *Mya arenaria* invaded the Baltic Sea likely by ships from North America (Strasser 1999) or even earlier by the Vikings (Petersen et al. 1992; Behrends et al. 2005). However, because this stage does not indicate natural environmental, climatic, or hydrographical changes of the Baltic Sea and is considered to be the result of anthropogenic activities, most recent authors have not discussed this stage (Andrén 1999; Brenner 2005).

2.1. Baltic Ice Lake

Glacial melt waters formed the BIL during the deglaciation of the Late Weichselian ice sheet. The stage of BIL covers the time period of ca 16 000 to 11 700 cal BP (Andrén et al. 2011).

The northern shore of the BIL was formed by the retreating ice sheet margin. The ice sheet re-advanced and retreated in several steps, resulting in five stages of the BIL in the eastern Baltic. These stages include A1, ca 13 800 (the Pandivere–Neva stade); A2, 13 300–13 200 (the Palivere stade), as revised by Saarse et al. (2012a, 2012b); BI, 12 250 (the standstill of ice margin at Salpausselkä I); BII, 12 000 (the retreat of the ice margin to the Salpausselkä II); and BIII, 11 590 cal BP (the ice margin standstill at the Salpausselkä II; Saarnisto & Saarinen 2001; Johansson et al. 2011). The shoreline formations of these BIL stages are traceable in the eastern Baltic. The highest shoreline of A1 is recorded at 90 m above sea level (a.s.l.) in the northern Estonia and at 50 m a.s.l. in the northwestern Latvia (Rosentau et al. 2009; Vassiljev et al. 2011). According to Zelčs & Markots (2004), however, the shoreline of BIL in the territory of Latvia reached up to 55 m a.s.l. Just prior to the Billingen drainage, the BIII shoreline in southern Finland was at 150 m a.s.l., 65 m a.s.l. in western and northern Estonia, and 40 m a.s.l. in northwestern Latvia (Vassiljev et al. 2011). In general, the shore displacement of BIL in eastern Baltic owing to land uplift was regressive.

Aquatic primary production in BIL was low (Andrén et al. 2011) because the melt water from the decaying ice margin contained abundant clastic material, and suspended particles from large proglacial rivers made the water turbid and impenetrable for sunlight. Owing to inappropriate living conditions, sediments are mostly barren of diatom frustules and other fossils. Moreover, Heinsalu (2001) suggested that the very high sedimentation rates during the deglaciation might have caused a dilution effect in the diatoms.

The typical feature of the sediment sequences of BIL is annually laminated varved clays, which formed in the vicinity of the receding ice margin, whereas homogenous clays were deposited further to the south, away from the ice margin. In the eastern Baltic, varved clays have been studied by Kuršs & Stinkule (1966), Danilans (1973), Veinbergs et al. (1980), Hang (1997), Strömberg (2005), and Hang et al. (2007).

2.2. Yoldia Sea

The last drainage of BIL through south–central Sweden, known as Billingen drainage, at $11\,690 \pm 10$ varve years BP (Andrén et al. 2002) or $11\,653 \pm 99$ cal BP (Rasmussen et al. 2006) marks the beginning of the YS. This phase of slightly brackish-water Baltic Sea lasted ca 1000 years from 11 700 to 10 700 cal BP (Andrén et al. 2011). This stage was named after the arctic bivalve *Portlandia (Yoldia) arctica*. As the water level suddenly lowered approximately 25 m to the ocean level in one–two years (Andrén et al. 2002; Jakobsson et al. 2007; Wohlfarth et al. 2008), the landscape in the coastal area changed dramatically. Vast territories were exposed to aeolian and fluvial processes, such as the formation of dunes and new river networks, river-bed incision, and paludification. During the subsequent 1000 years, the water level of the YS was in balance with the ocean and was quite stable (Rosentau et al. 2011) although the sea level regressed in the northern Baltic Sea basin primarily because of land uplift. A low water level during the YS stage is indicated by submerged ancient tree stumps in the southern part of the Baltic Sea (Björck 1995; Bitinas et al. 2003; Björck et al. 2008; Uścińowicz et al. 2011), as well as buried peat layers (Bennike et al. 2000; Saarse et al. 2006) and paleosoils (Berglund et al. 2005). All of the aforementioned elements serve as index points for reconstructing the sea-level curve.

Several studies have been conducted on the salinity and other environmental proxies in the YS by Wastegård et al. (1995), Heinsalu (2000), Andrén et al. (2002), Brenner (2005), and Heinsalu & Veski (2007). According to salinity proxies, the history of the YS is divided into three phases: freshwater, brackish, and freshwater. During the first 200–300 years in the YS, freshwater conditions prevailed because the increased meltwater supply and large river input, which provided a strong outflow of freshwater to prevent saline water inflow into the basin through the narrow Närke Strait (Heinsalu 2001). The saline water inflow occurred simultaneously with the Preboreal oscillation at 11 300–11 150 cal BP (Björck et al. 1997; Heinsalu 2001). Deterioration in the climate decreased the melting of the Weichselian ice sheet, thereby hampering the freshwater supply (Björck et al. 1996) and promoting saline water intrusion over the threshold. In offshore areas, a stratified water column developed with saline water at the bottom and turbid freshwater on top owing to density differences. The salinity of the bottom waters close to the inflow area reached 10‰ (Wastegård 1997), whereas freshwater conditions remained in the offshore surface water settings of the Gulf of Finland (Heinsalu et al. 2000). In the coastal areas, brackish water likely was transported up to the surface by upwelling processes (Heinsalu 2001).

According to diatom taxa, the salinity in the coastal area of northern Estonia was similar to or slightly lower than modern levels (Heinsalu 2001), where the surface salinity varies from 5–7‰ in the western part to about 0–3‰ in the eastern part (Alenius et al. 1998). The final phase of the YS is characterised by freshwater conditions again because the climate repeatedly ameliorated and triggered ice-sheet melting to cause enhanced freshwater supply to the Baltic.

The shoreline of the YS is characterised as regressive due to rapid land uplift. The maximum height of the shoreline in northern Estonia was approximately 46–47 m a.s.l. (Künnapuu 1959); the water level dropped to at least 24 m a.s.l. by the end of YS (Heinsalu & Veski 2007). The highest shoreline on the Island of Hiiumaa, at 60–61 m (Kessel & Raukas 1967), is assumed to be a formation of the YS. It is rather difficult to trace the YS shoreline along the Latvian coast because no obvious evidence has been reported thus far. According to studies from the second half of the 20th century, the sea level in the southern part of the Gulf of Riga was at least 29 m b.s.l. (Berzin 1967), as well as 15 m b.s.l. (Kovalenko & Juskevics 1987). If the water level of BIL just prior to drainage reached up to 16 m a.s.l. in the coastal area of Latvia (Vassiljev & Saarse 2013), then the maximum level of the YS should have been approximately 9 m b.s.l. However, the minimum level is uncertain owing to a lack of data.

2.3. Ancylus Lake

The connection of the Baltic Sea basin to the ocean through Närke Strait shallowed and finally closed owing to sustained land uplift. The water level within the Baltic Sea basin began to rise, and the next freshwater stage of the Baltic Sea, AL, was formed from 10 700 to 9800 cal BP (Andrén et al. 2011). This stage was named after gastropod mollusc *Ancylus fluviatilis*, which is widespread in the AL sediments. The AL was fed by input of the last remnants of the melting ice sheet, and rivers flowing through recently deglaciated the drainage area with fairly pristine soils. Hence it was a well-mixed oxygenated water body with low nutrient input (Andrén et al. 2011). Moreover, the high abundance of diatoms at the onset of the AL stage indicates that a significant amount of organic carbon was produced in the water basin. In shallower waters during the AL transgression, however, the organic carbon might to a certain extent originate from eroded and redeposited shore deposits (Andrén et al. 2000a).

The time span for the AL transgression is estimated to be around 500 years (Andrén et al. 2011). The maximum of the transgression occurred around 10 200 cal BP (Saarse et al. 2003; Berglund et al. 2005; Björck et al. 2008; Wohlfarth et al. 2008). Owing to its transgressive character, the shoreline of the AL in Estonia was displaced towards the mainland. On the Kõpu Peninsula in the Island of Hiiumaa, the highest shoreline of AL reached up to 45 m a.s.l.; that in northern Estonia was 35 m (Kents 1939) or even 37.4 m (Künnapuu 1959). In Latvia, some parts of the AL shoreline are traceable in northern Courland (Kurzeme), as proposed by Grinbergs (1957), slightly above the LitS shoreline. On the contrary, Meirons cited in Veinbergs (1996) considered that the AL shoreline is at the same

altitude or slightly above the present sea level. In the Gulf of Riga, the AL shoreline is beneath the present sea level; Veinbergs (1996) and Kovalenko and Juskevics (1987) reported b.s.l values of 20 m and 5–6 m, respectively.

After the AL reached its culmination, the dammed waters found a new outlet through the Danish straits (Björck 1995). Discussions among scientists remain about the length and amplitude of the AL regression, the location of actual threshold, the role of incised Dana River, and other possible outlets to the ocean. Although there are difficulties in determining the low AL level because of partially preserved or buried deposits and relief forms under the transgressive deposits of the LitS (Raukas 1999), the reconstructed shoreline displacement curves suggest rapid regression with an amplitude of approximately 20 m in the central, eastern, and southern Baltic (Saarse et al. 1997; Raukas 1999; Saarse et al. 2000; Berglund et al. 2005; Veski et al. 2005; Wohlfarth et al. 2008). Research by Björck (1995) and Novak & Björck (1998) implies that the regression of AL occurred during a very short time period of two–three years. Other studies reject this concept. Jensen et al. (1999), Bennike et al. (2000, 2004), and Lemke et al. (2001) deny the rapid drainage of the AL because there is no indication of a rapid lowering of the AL level in the early Holocene deposits in the Great Belt. Moreover, the altitudes of possible thresholds do not correlate with reconstructed amplitude of the AL water level drop. Björck et al. (2008) attempted to find a compromise between the model of sudden and massive drainage of the AL and the concept of a long and calm regression. As a result, they suggested that the final drainage might have occurred through the German–Danish area, but occasional marine water inflows into the Initial LitS turned out through the Dana river system (Björck et al. 2008). Either way, the location of the AL threshold remains undetermined. Bennike et al. (2004) suggested that the water level drop of the AL could have been caused by subsidence of the critical threshold; however, Lambeck (1999) reported that only a few metres of subsidence could be expected over the relevant time period.

2.4. Litorina Sea

The transition from the AL to LitS occurred at around 9800 cal BP (Andrén 1999) when the basin of Baltic Sea re-established its connection with the ocean via the Danish Straits. This stage was named after gastropod mollusc *Littorina littorea*. The onset of the LitS can be determined through the rapid increase in organic material (OM) in the sediments caused by enhanced primary production as a response to climate warming and upwelling of nutrient-enriched bottom waters during marine water inflows (Andrén et al. 2000a).

A period of a slightly brackish water phase between 9800 and 8500 (8000) cal BP is known as the Initial LitS (Andrén et al. 2000b; Lampe 2005; Andrén et al. 2011) or Early LitS (Risberg 1991; Berglund et al. 2005). However, some authors denominate this phase as the Mastogloia Sea (Alhonen 1971; Hyvärinen et al. 1992; Wastegård et al. 1998; Jensen et al. 1999; Miettinen 2002; Tikkanen & Oksanen 2002) after the diatom species *Mastogloia* spp. commonly found in the coastal deposits. The Initial LitS, owing to interplay between eustatic sea-

level rise and glacio–isostasy, was time-transgressive and ended around 8500 cal BP when the brackish environment was established throughout the Baltic Sea basin (Andrén 1999). At the beginning of the Initial LitS, salinity in the southern part of Baltic Sea basin increased slowly because of weak and intermittent seawater influxes (Andrén et al. 2000a; Berglund et al. 2005).

Owing to a eustatic sea level (ESL) rise between 8500 and 8000 cal BP, the Danish straits widened, providing more comprehensive saline water influxes from the North Sea into the Baltic Sea basin (Westman et al. 1999). Gradually, brackish conditions were established throughout the Baltic Sea basin.

There is no doubt that the LitS was transgressive in the southern part of the Baltic Sea basin; however, the number of transgressions is debated. Yu et al. (2003, 2004), Sandgren et al. (2004), and Berglund et al. (2005) support the concept of a multi-transgression pattern across the southern Baltic Sea with rapid ESL rise at 7600 cal BP (Yu et al. 2007). These authors reported that several minor transgressions occurred synchronously between 8000 and 5500 cal BP in northwestern Russia and southeastern Sweden. Three LitS transgressions are shown in the shoreline displacement curve of the Lithuanian coast (Gelumauskaitė 2009; Damušytė 2011). A twofold transgression was recognised in Russia (Miettinen et al. 2007), Latvia (Grinbergs 1957; Ulsts 1957; Veinbergs 1996; Eberhards 2006), and northern Germany (Lampe 2002). The recent studies from Finland (Seppä et al. 2000; Miettinen 2002, 2004) and Estonia (Veski et al. 2005; Saarse et al. 2009a; 2009b; Rosentau et al. 2013) confirm a single large-scale LitS transgression at 7800–7400 cal BP followed by a linear regression with no remarkable sea-level fluctuation. Uścińowicz (2006) proposed one long-lasting LitS transgression along the Polish coastline similar to that in the southern part of the Baltic Sea, where a gradual sea-level rise has been established since the Baltic Sea basin became permanently connected to the ocean.

The culmination of the LitS transgression in areas with slow land uplift occurred later than that in areas of rapid land uplift (Miettinen 2002). The highest LitS limit in Hiiumaa is at 25 m a.s.l.; that in northwestern Estonia is 21–22 m a.s.l. (Kessel & Raukas 1979). The LitS transgression in northern Estonia reached approximately 22 m a.s.l., and the magnitude was about 4–5 m (Saarse et al. 2009a). Grinbergs (1957), Ulst (1957), and Veinbergs (1996) reported that the highest shoreline of the LitS in Latvia is at 12–13 m a.s.l. in northern Courland and 5 m a.s.l. in the southern part of the Gulf of Riga (Veinbergs 1996).

2.5. Limnea Sea

The LimS, or Post-Litorina, represents the youngest stage of the Baltic Sea from 4500 cal BP until the present (Kessel & Raukas 1979). The term ‘Limnea Sea’, named after the mollusc *Lymnaea ovata* f. *baltica*, is used commonly in Estonia and Finland (Hyvärinen et al. 1992; Seppä et al. 2000; Tikkanen & Oksanen 2002; Miettinen et al. 2004; Saarse et al. 2010), although ‘Post-Litorina’ is more common in Sweden, Poland, and Germany (Andrén 1999; Jankowska et al. 2005; Witak et al. 2005; Miltner et al. 2005). The transition from LitS to LimS is marked

by a decrease in salinity and hence the disappearance of most marine components such as certain silicoflagellates, dinoflagellates, and *Chaetoceros mitra* resting spores (Andrén 1999; Brenner 2005).

The reconstructed sea level curves around the Gulf of Bothnia (Lindén et al. 2006; Widerlund & Andersson 2011) and the Gulf of Finland (Miettinen 2002, 2004; Saarnisto 2012) show a linear, continuous trend of sea level decrease. Moreover, Seppä & Tikkanen (1998) stated that the sea level of the Baltic Sea, at least in the northern part with rapid land uplift, has continued decrease uninterrupted during the last millennia. At the same time, in subsiding areas such as the coasts of Denmark, Germany, Poland, and Lithuania, transgression of the sea is still occurring. This fact is evidenced by the occurrence of submerged tree stumps not deeper than 1 m b.s.l. in the southwestern part of the Baltic Sea (Lampe 2005).

Bitinas et al. (2002) reported that the LimS stage in Lithuania began with a transgression about 4000 years ago, when water level rose up to 1–2 m a.s.l. Nevertheless, evidence for such a high water level has not been found in the coastal peat lands along the southern coast of the Baltic Sea. Most likely, the traces of a surge have been confused with evidence of temporal sea-level fluctuations (Lampe 2002). Yu (2003) discussed the substantial regression in south–central Sweden at about 3000 cal BP, and Damušytė (2011) reported that the LimS level in Lithuania began to fall at 2400–2200 cal BP until reaching the present level. However, the data from Poland presented by Uścińowicz (2006) demonstrate slow sea-level rise over the last 3000 years with an average rate of 0.3–0.25 mm yr⁻¹; hence, the present sea level was reached in that region about 200–100 years ago. Moreover, Lampe (2002) discussed several minor transgressions and regressions of the LimS in the southern Baltic Sea basin, mentioning a remarkable sea-level rise at about 1000–600 cal BP. Contradictory data imply that sea-level changes are most likely the result of short-term severing events such as breached barriers by storm surges and salt water inflows into coastal water basins (Uścińowicz 2006).

3. DIATOMS IN THE ISOLATION BASINS

Diatoms are a group of microscopic algae abundant in almost all aquatic habitats (Stoermer & Smol 1999). Their distribution depends on different factors such as temperature, water transparency, turbulence, ice cover length, pH, nutrients, and salinity (Battarbee et al. 2001). Due to the ability of diatoms to quickly respond to changes in physical, chemical, or biological conditions and their good preservation in sediments, palaeolimnologists widely apply diatom analysis as an indicator of environmental change. Diatom analysis has often been used in isolation basin studies to reconstruct past sea-level fluctuations in areas affected by glacio–isostasy such as Sweden (Hedenström & Risberg 1999; Westman & Hedenström 2002; Yu et al. 2004; Risberg et al. 2005), Finland (Seppä et al. 1998; Seppä & Tikkanen 2000; Eronen 2001; Miettinen 2004; Ojala et al. 2005), Norway (Corner et al. 1999; Balascio et al. 2011, Barnett et al. 2015), Estonia (Heinsalu & Veski 2007; Saarse et al. 2000, 2009a, 2009b), Russia (Heinsalu et al. 2000; Sandgren et al. 2004), Scotland (Shennan et al. 1996, 2000), Iceland (Saher et al. 2015), and Greenland (Long et al. 2011).

Isolation basins are natural topographic depressions that once were part of a larger water basin or were influenced once or various times by the sea transgression in their history owing to eustatic sea-level rise or isostatic rebound (Long et al. 2011). The term ‘isolation basin’ is in some studies referred to as ‘ancient lagoon’ (e.g. Yu et al. 2004; Saarse et al. 2009a, 2009b). The isolation contact in the sediment sequence can be recognised by changes in diatom assemblage. For example, marine diatom replacement by freshwater diatoms indicates a shift from a saline to freshwater environment.

In this thesis, the isolation basins of the various Baltic Sea stages are investigated. Hence, it is useful to examine the most common diatom assemblages in every stage of development of the Baltic Sea.

Diatom taxa (Fig. 1) typical of particular development stages of the Baltic Sea are listed in Snoeijs (1999). Sediments of the early stages of BIL are usually barren of diatom frustules. The most common diatoms in the sediments of the BIL are freshwater planktonic *Aulacoseira islandica*, *A. alpigena*, *Stephanodiscus rotula*, and *Tabellaria fenestrata*. The YS was slightly brackish, containing a mix of marine water and cold fresh water. Therefore, diatoms that tolerate minor salinity such as *Actinocyclus octonarius*, *Diploneis didyma*, *D. interrupta*, *D. smithii*, *Thalassiosira baltica*, *Tryblionella navicularis*, and *T. punctata* prevailed. The AL diatom flora is characteristic with planktonic freshwater taxa *Aulacoseira islandica* and *Stephanodiscus neoastraea* and periphytic taxa such as *Cocconeis disculus*, *Cymatopleura elliptica*, *Diploneis domblittensis*, *Ellerbeckia arenaria*, *Encyonema prostratum*, *Gyrosigma attenuatum*, and *Navicula jentzschii*. When marine water again started to flow into the Baltic Sea basin, AL was replaced by the LitS. The most common diatom taxa during the Initial LitS stage were *Diploneis smithii*, *Epithemia turgida*, *E. turgida* var. *westermanni*, *Mastogloia braunii*, *M. elliptica*, *M. pumila*, *M. smithii*, *N. peregrina*, and *Rhoicosphenia curvata*. As the environment gradually

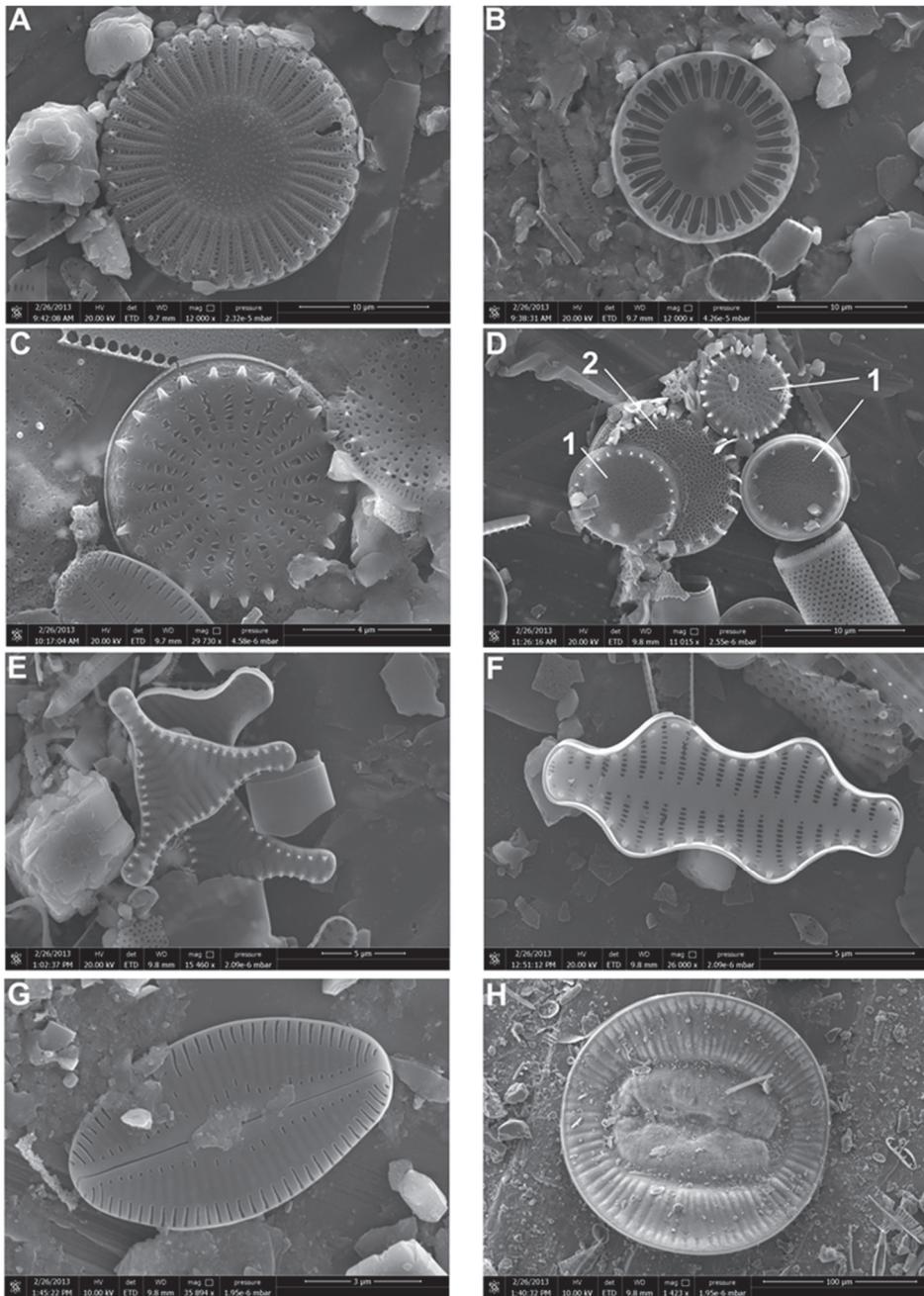


Figure 1. Diatoms from isolation basins. **A** – external valve of *Cyclotella meneghiniana*; **B** – internal valve of *C. meneghiniana*; **C** – *Stephanodiscus parvus* (A, B, C from Lake Harku core at depth 250 cm); **D** – **1** *S. parvus* (external and internal valve), **2** – *S. hantzschii*; **E** - *Staurosira construens* var. *exigua* (D, E from Harku core at depth 180 cm); **F** – *S. construens* var. *binodis* (Lake Harku core at depth 320 cm); **G** – *Achnanthes fagedii*; **H** – *Campylodiscus clypeus* (G, H from Lake Klooga core at depth 420 cm). Photos made by Marianne Ahlbom from Department of Geological Sciences at Stockholm University.

became more brackish, up to 10–15‰ (Gustafsson & Westman 2002; Emeis et al. 2003), diatom taxa with higher salinity optima appeared, such as *Campylodiscus clypeus* (Fig. 1H), *Cocconeis scutellum*, *Diploneis didyma*, *D. interrupta*, *Chaetoceros diadema*, *C. mitra*, and *Hyalodiscus scoticus*. The subfossil *Achnanthes fagedii* (Fig. 1G) is a typical species of LitS (Snoeijs & Kasperovičienė 1996; Witkowski et al. 2000).

Because isolation commonly occurs gradually, it is possible to distinguish a transition phase in the sediment sequence. The length of the transition varies from very short or almost imperceptible in areas with rather high land uplift rate to distinct, thick sediment layers where gradual increases of OM and mass occurrences of small-sized fragilarioid taxa are observed. An extended transition phase, or an isolation phase, usually occurs in areas with relatively flat topography and a slow apparent land uplift rate.

Discussions about mass occurrences of small-sized benthic fragilarioid diatoms often appear in articles about diatoms in isolation basins. Denys (1990) suggested that mass occurrence of *Fragilaria* spp. is related to environmental instability. This group is considered to be pioneer taxa, which are therefore able to adapt to changes faster and more effectively than other diatoms (Yu et al. 2004). The predominance of *Fragilaria* spp. in the isolation basins just prior to, during, or after the actual isolation event was examined in a seminal paper by Stabell (1985). The same pattern has been observed in other studies (e.g. Heinsalu et al. 2000; Seppä et al. 2000; Risberg et al. 2005; Miettinen et al. 2007; Saarse et al. 2009a) which address diatoms in the isolation basins. Mass occurrences of small-sized fragilarioid taxa can be explained by their ability to outcompete other diatoms in unstable environment marked by changes in salinity, pH, temperature, and nutrient enrichment. The pH rise is caused by carbonate-enriched Quaternary deposits or Silurian and Ordovician bedrock. Another frequently occurring feature of changes in nutrient content and salinity in isolation basins is the appearance of *Chaetoceros* spp. resting spores (Andrén et al. 2000b; Risberg et al. 2005). Their mass occurrence can be interpreted as either the result of increased access to nutrients or a sign of deteriorated living conditions (Snoeijs 1999). The transition phase is also indicated by so-called *Clypeus* flora such as *Campylodiscus clypeus*, *Anomoeoneis sphaerophora*, *Tryblionella circumscuta*, *Nitzschia scalaris*, and *Surirella striatula*, which thrive in the brackish, nutrient-rich environment typical of shallow sea bays or lagoons (Snoeijs 1999; Risberg et al. 2005). In addition, *Surirella biseriata* is common in shallow freshwater lakes that are isolated from the sea.

After the isolation event, the diatom composition developed towards freshwater small-lake flora. Common species found in the isolated basin are planktonic *Aulacoseira ambigua*, periphytic *Navicula schoenfeldii*, *Cymbella diluviana* (Heinsalu 2000), *Cymbella ehrenbergii*, *Gomphonema angustatum*, *Mastogloia smithii* var. *lacustris*, and *Navicula oblonga* (Saarse et al. 2009a). Marine and brackish diatom appearances after the isolation event can indicate temporary sea water inflow into the lake or resedimentation. Therefore, it is essential to pay attention to the preservation of diatom frustules. If they are

broken and physically eroded and include only some marine/brackish species among freshwater diatoms, they are likely reworked from older sediments. To unravel the processes of redeposition, the entire microfossil complex should be considered (Liivrand 1999).

4. MATERIAL AND METHODS

4.1. Study area

The study area covers the eastern part of the Baltic Sea that encompasses northern and northwestern Estonia (Papers I–III), the Island of Hiiumaa (Paper IV), and the central part of Latvia, particularly the coastal area of the southern part of the Gulf of Riga (Paper V; Fig. 2A).

The coastline of the Baltic Sea is greatly modified due to glacio–isostatic adjustment. In areas where land uplift is higher than sea-level rise, such as Estonia, new land such as islands and peninsulas have appeared over time, forming a highly indented coastline. Conversely, the coastline along Latvia, where sea-level rise is equal to or exceeds the land uplift, is straightened. The coastal area along the eastern Baltic Sea is characteristic of relatively flat topography with gentle sloping towards the sea, resulting in extensive paludification around the investigated lake basins. Two sites included in this study, Loopsoo and Kõivasoo bogs, are already overgrown lakes.

4.1.1. Northern and northwestern Estonia

Five of the studied sites are located in northern and northwestern Estonia. Lakes Harku, Lohja, and Käsma (Fig. 2B; Table 1) are situated in the North Estonian coastal plain that was once part of a former basin of the Gulf of Finland. Hence, it is relatively flat and rich in rather small and shallow lakes and paludified areas; most of Quaternary sediments and relief forms are of marine origin that evolved since the YS stage. Currently, the coastal plain is a narrow strip between the periglacial depression of the Gulf of Finland and the Baltic Klint, representing a coastal escarpment that was eroded in the sedimentary rocks of the Cambrian and the Ordovician that overlie the crystalline basement.

Lake Harku is a medium-sized, shallow hypereutrophic lake located at the back of Kakumäe Klint Bay on the western border of Tallinn. The lake shores are flat and partly paludified and are covered by meadows, pastures, and a rim of *Alnus* and *Salix*. Its catchment is rather densely settled, resulting in increasing anthropogenic stress. **Lake Lohja** is a shallow, dark-coloured soft-water lake located in the southwestern corner of the Pärисpea Peninsula. Before its isolation, its basin was located at the bottom of Hara Bay, which was opened on the northwestern side of the lake. Currently, the lake is surrounded in the north and west by beach ridges and dunes evolved during the LimS regression. Similar to Lake Lohja, **Lake Käsma** is a small, dark-coloured soft-water lake surrounded by beach ridges in the west and a flat, slightly paludified marine plain in the east covered by various types of forest.

Lakes Tänavjärv and Klooga (Table 1) are situated in northwestern Estonia within the low and flat topography formed on Ordovician limestone. Hence, dominating landscapes are present with limestone bedrock knolls, escarpments, and calcareous soils with characteristic calciphilous vegetation known as alvars.

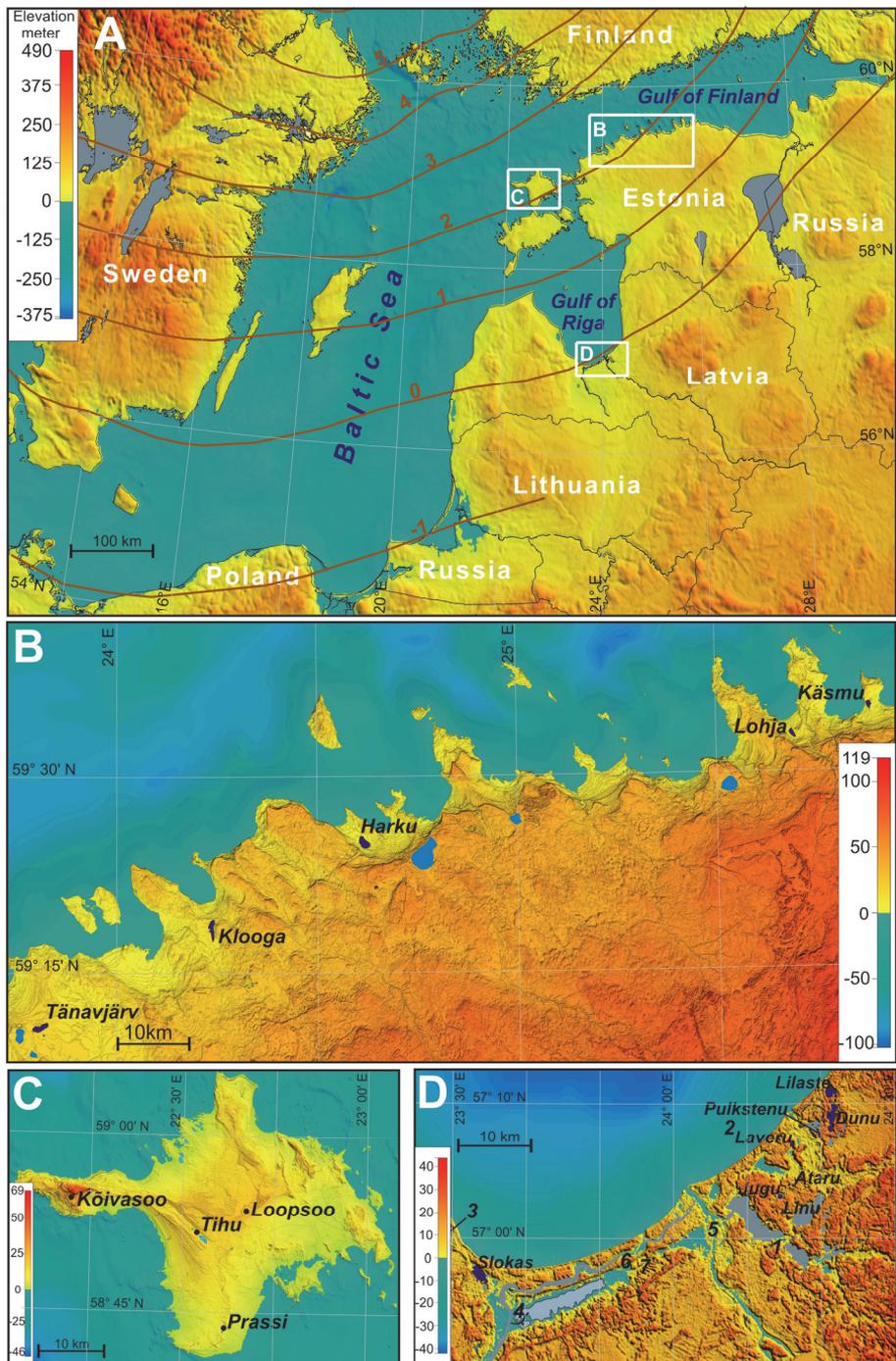


Figure 2. **A** – Overview map of the study area, brown lines show apparent uplift in mm yr⁻¹ (Ekman 1996). Digital terrain model of: **B** – the northern and northwestern Estonia, **C** – the Island of Hiiumaa, **D** – the coastal area of southern coast of the Gulf of Riga. Additional sites mentioned in discussion: **1** – Babelitis; **2** – Sand deposits ‘Vecdaugava’; **3** – Kanieris; **4** – Babites; **5** – Daugava; **6** – Priedaine; **7** – Slepere.

The Quaternary till is covered by rather thin marine and aeolian sands except in buried valleys, where it can reach up to 37–46 m (Kadastik & Ploom 2000). **Lake Tänävjärv** is a shallow, semi-dystrophic soft-water lake located in the middle of Tänävjärv Bog. Therefore, the shores of the lake are peaty; in the west and east, however, they are partly sandy. The catchment is forested by boreal tree species, mostly *Pinus sylvestris*, and is frequently subject to forest fires (Kangur 2005). **Lake Klooga** is a shallow drainage lake largely overgrown by emergent aquatic macrophytes. Its bedrock consists mostly of limestone; thus, small brooks and bottom springs carry calcareous water to the lake and promote the precipitation of lacustrine lime. The catchment area of the lake is covered by a pine forest to the north and fields and meadows to the east; to the south and west, it is paludified and forested.

Table 1. The description of study sites in the north and northwestern Estonia.

Site	Coordinates	Area (km ²)	Altitude (m a.s.l.)	Water depth (m)		Threshold (m a.s.l.)
				Mean	Max	
Lake Tänävjärv	59°10'58" N 23°48'40" E	1.388	18.4	2.0	3.0	17.4
Lake Klooga	59°21'32" N 24°15'00" E	1.314	11.8	1.9	3.7	11.8
Lake Harku	59°25'00" N 24°37'06" E	1.633	1.2	1.7	2.3	ca 2.5
Lake Lohja	59°32'55" N 25°41'30" E	0.568	5.5	2.2	3.7	6.0
Lake Käsmu	59°34'49" N 25°53'01" E	0.485	3.9	2.0	3.3	4.5

4.1.2. Island of Hiiumaa

The Island of Hiiumaa is part of the West Estonian archipelago located in the eastern part of the Baltic Sea (Fig. 2C). It is rather young in its present shape and was formed during the Holocene starting with the YS stage, when the highest point, Tornimägi (68 m a.s.l.), emerged from the sea around 11 000 years ago (Paper IV). The geology is similar to the aforementioned stratigraphy of northern Estonia. The basement of Hiiumaa is composed of Precambrian crystalline rocks covered by Cambrian sandstones, siltstones, and clays overlain by Ordovician and Silurian limestones and dolostones. Most of the bedrock is covered by Quaternary sediments such as till, glaciofluvial gravel, marine sand and silt, and aeolian sand. Emerging from the sea, the initial glacial landscape morphology was significantly altered by sea waves, drift ice, and wind erosion owing to glacio–isostatic adjustment of the earth’s crust (Raukas et al. 2009). This process has created beach ridge systems, fan-like bars, sandy terraces, and spits sporadically capped by dunes (Eltermann 1993a, 1993b). The nature, geology and history of the island have been studied by several authors (Kessel & Raukas 1967,

1979; Sepp 1974; Raukas & Ratas 1996; Hang & Kokovkin 1999; Königsson et al. 1998; Saarse et al. 2000).

Three sites were chosen for detailed sea-level studies: Loopsoo Bog, Lake Tihu Keskmine, and Lake Prassi. To obtain more accurate index points for sea-level reconstructions, earlier published data of Kõivasoo Bog (Fig. 2C; Table 2) appearing in Königsson et al. (1998) and Saarse et al. (2000) were added.

Kõivasoo, an ancient LitS lagoon on the Kõpu Peninsula, is a small raised bog that lies near the LitS limit between two series of scarps at 27.6 and 23.8 m a.s.l. It contains lagoonal, limnic, and terrestrial deposits (Saarse et al. 2000). **Loopsoo** is an ancient, overgrown lagoon included in the large swamp area in the central part of Hiiumaa. It is surrounded by bow-shaped beach ridges, a thin woody rim, and field patches. Orru (1995) studied the thickness, distribution, and properties of the Loopsoo peat. **Lake Tihu Keskmine**, hereafter referred to as **Tihu**, is a small, elongated, shallow, semidystrophic lake in the western part of Hiiumaa. This area is characteristic with brownish till with erratic clasts and widely distributed marine sand, and the topography is broken by many beach ridges, spits, and fan-like bars (Ratas 1976; Eltermann 1993b). The lake shores are overgrown and are flanked by ridges and dunes; the catchment is paludified and is mostly forested by pines. **Lake Prassi** is a small lake in southern part of Hiiumaa that is mainly overgrown by reeds and surrounded by forested mire.

Table 2. The description of study sites on the Island of Hiiumaa.

Site	Coordinates	Area (km ²)	Altitude (m a.s.l.)	Water depth (m)		Threshold (m a.s.l.)
				Mean	Max	
Kõivasoo Bog	58°54'30" N 22°11'55" E	1.836	27.5	-	-	27.0
Loopsoo Bog	58°53'40" N 22°40'20" E	2.741	21.5	-	-	21.0
Lake Tihu Keskmine	58°51'47" N 22°32'21" E	0.027	14.5	-	0.6	14.4
Lake Prassi	58°43'41" N 22°36'59" E	0.075	7.2	-	-	6.8

4.1.3. Southern coast of Gulf of Riga

The area studied in the central part of Latvia covers the Rigavas sandy plain, which is part of the coastal lowland along the southern coast of the Gulf of Riga. In total, eight lakes were cored (Fig. 2D; Table 3). Lakes Lilaste, Dunu, and Pulkstenu are located in the northeastern Rigavas Plain; lakes Ataru, Linu, Laveru, and Jugu are in the central part of the plain located between two large rivers, Gauja and Daugava; and Lake Slokas is distant from other sites, at the southeastern margin of the Rigavas Plain (Fig. 2D).

The coastal lowland is an abrasional and accumulation plain originating from BIL and is the latest post-glacial formation presently in continuous development.

The inland border of the coastal lowland is considered to be the highest shoreline of BIL. Flat terrain and dune barriers along the sea coast have promoted mire and bog development that are common at the southern coast of the Gulf of Riga.

Rigavas Plain is an ancient dune and sandbar system that interchanges with low plains and wetlands covered by pine forest. The average height of the territory is 9–14 m a.s.l., and the altitude of the dunes reaches up to 16–20 m. The depressions between the dune ridges are filled by many shallow, partly overgrown lakes of various sizes. The surface of the plain is crossed by wide (1–4 km) but shallow (5–10 m) valleys of three large rivers: the Lielupe, Daugava, and Gauja.

Compared with the study area in Estonia, the crystalline basement in central Latvia is deeper, at >1000 m b.s.l. The crystalline rocks are covered by Cambrian sandstones, silts, and clays; Ordovician and Silurian carbonate rocks and clays; and Devonian sandstones and dolomites. The thickness of the pre-Quaternary deposits ranges from 300 m to 1000 m. The Quaternary cover consists of Weichselian tills, glaciolacustrine and marine silts, sands, and aeolian deposits. The thickness of the Quaternary cover of the southern coast of Gulf of Riga ranges from 10 m to 40 m in the northeast and 40 m to 80 m in the area between the Gauja and Daugava rivers; proceeding towards the southwestern margin, it decreases up to <10 m (Kovalenko & Juskevics 1987).

Lakes Lilaste and Dunu, once formed a larger waterbody and lagoon of BIL, are presently still connected with a 900-m long overgrown channel. Both Lilaste and Dunu are drainage lakes, with inflow from Melnupe and Puska rivers and outflow via the Lilaste River to the Gulf of Riga.

Lake Pulkstenu is a small lake with steep shores located in a depression between high dunes covered by pine forest. **Lakes Ataru, Laveru, Jugu, and Linu** are small lakes located in a plain between the Gauja and Daugava rivers, which could be an accumulation delta combined with distributaries of the Daugava and Gauja rivers (Saule-Sleinis 1960).

Lake Slokas is a large, shallow coastal lake situated in Kemeru National Park. The lake shores are overgrown by reeds, however some parts of the shores are devoid of vegetation. The bottom of the lake is situated on Devonian dolomite bedrock. This area is well known for its sulphur springs, which are formed by organic, rich bog water seepage through the Upper Devonian gypsum, which forms hydrogen sulphide–calcium sulphate brines.

Table 3. The description of study sites in the Rigavas Plain, the southern coast of the Gulf of Riga.

Site	Coordinates	Area (km ²)	Altitude (m a.s.l.)	Water depth (m)		Threshold (m a.s.l.)
				Mean	Max	
Lake Lilaste	57°10'44" N 24°21'06" E	1.836	0.5	2	3.2	0.3
Lake Dunu	57°08'58" N 24°21'28" E	2.741	0.7	1.6	3.3	0.3

Lake Ataru	57°03'52" N 24°16'06" E	0.136	3.2	2.2	3.3	3.8
Lake Laveru	57°06'55" N 24°16'06" E	0.21	1.2	0.9	2.3	ca 1.5
Lake Pulkstenu	57°07'47" N 24°19'43" E	0.038	1.9	2.4	5	ca 4.5
Lake Jugu	57°03'13" N 24°14'21" E	0.355	4.1	0.8	3	ca 4.5
Lake Linu	57°02'00" N 24°14'41" E	0.06	3.3	0.3	1	ca 4.6
Lake Slokas	56°57'44" N 23°33'16" E	2.5	1.3	0.6	1.5	0.3–1.9

4.2. Methods

Sediment sampling

During field work expeditions in the winters of 2011–2013 in northern Estonia and central Latvia, sediment samples were obtained from the deepest parts of the basins through the ice by using a 10 cm in diameter, 1-m long Russian-type corer. In the case of Island of Hiiumaa, sediment samples were collected during the summer of 2012.

The unconsolidated sediments in the upper layers of lakes Harku and Lilaste were sampled by using a Willner-type sampler, and 2-cm slices were cut and placed into plastic bags. Core sections of 1 m in length were described in the field, photographed, sealed in plastic liners, and stored in a laboratory cold room.

Loss-on-ignition, magnetic susceptibility and grain size

The lithology of the cores was studied by using the loss-on-ignition (LOI) method following Heiri et al. (2001), and magnetic susceptibility (MS) following Nowaczyk (2001). To obtain additional information on the environment of the sedimentation processes in lake sediments, the grain size distribution was analysed by using a laser-scattering particle size distribution analyser (Horiba LA-950V2), and the samples were prepared following Vaasma (2008).

The OM content for continuous 1-cm and 2-cm-thick sediment samples was quantified by LOI at 525°C for 4 h, expressed as percentage of dry matter. The carbonate content was estimated in terms of the difference between LOI at 900°C for 2 h and 525 °C multiplied by 1.36. The ignition residue was estimated on the basis of mineral matter (MM) content.

Volume specific MS κ , expressed in SI units, was measured by using a Bartington MS2E high-resolution scanning sensor at 1-cm resolution on the cleaned sediment surface covered with a thin plastic film.

Diatom analysis

The samples for diatom analysis were prepared by using the standard method described in Battarbee et al. (2001). Sediment samples were digested in 30% hydrogen peroxide to remove OM, and a few drops of 10% hydrochloric acid were subsequently added to remove carbonates, metal salts, and oxides. Finally, to extract fine and coarse mineral particles, repeated decantation was applied. In the case of Lake Lilaste, a known quantity of synthetic microspheres was added to the diatom suspension to calculate the diatom accumulation rate (DAR) and *Chaetoceros* resting spore accumulation rate (AR).

A drop of the cleaned sub-sample was spread over the cover slip, dried overnight at room temperature, and mounted onto microscope slides with Naphrax medium. About 400–500 diatom valves were counted and were identified at the species level under oil immersion by using a Zeiss Axio Imager A1 microscope with differential interference contrast illumination at $\times 1000$ magnification. The diatoms were grouped in accordance with their salinity tolerance as marine/brackish, including marine and brackish-water forms occurring at salinities of 0.5‰ to 30‰; halophilous, including freshwater forms stimulated by low salinity; small-sized fragilarioid taxa with brackish water affinity; small-sized fragilarioid taxa preferring fresh water; indifferent taxa, including freshwater forms that can tolerate low salinity; and freshwater taxa. Habitat classification included planktonic, small-sized fragilarioid, and periphytic taxa. The diatom floras used for identification and ecological information were based on well-established sources (Krammer and Lange-Bertalot 1986, 1988, 1991a, 1991b; Snoeijs 1993; Snoeijs and Vilbaste 1994; Snoeijs and Potapova 1995; Snoeijs and Kasperovičienė 1996; Snoeijs and Balashova 1998; Witkowski 1994; Witkowski et al. 2000).

Dating

For all studied sites, dating was conducted on 41 samples by Accelerator Mass Spectrometry (AMS; Poznan Radiocarbon Laboratory, laboratory code: Poz) and 9 by conventional ^{14}C methods (Institute of Geology, Tallinn University of Technology, laboratory code: Tln; Göran Possnert Tandem Laboratory, University of Uppsala, laboratory code: Ua; University of Tartu, laboratory code: Ta; Table 4).

Age-depth models in papers III and V were produced by using the IntCal13 calibration dataset (Reimer et al. 2013) and the OxCal 4.2.4 deposition model (Bronk Ramsey 2009, Bronk Ramsey & Lee 2013), where the ^{14}C dates were combined with lithological boundaries. Additionally, ^{210}Pb dating at Lake Harku and spheroidal fly ash particles at Lake Lilaste were used to date the upper loose lake sediments. The peak in spheroidal fly ash particles at Lake Lilaste at a core depth of 218 cm occurred in model year 1982 ± 10 AD, which corresponds to the maximum air pollution according to the Latvenergo AS report on emissions in the atmosphere. The radiocarbon ages were calibrated at a 95.4% probability

range, and the weighted averages before the present were used (cal BP, 0 = 1950 AD).

Table 4. Radiocarbon dates calibrated at 95.4% probability range from studied lakes sequences.

Study site, (altitude, m a.s.l.)	Depth, cm	¹⁴ C date, BP	Calibrated age, BP (weighted average)	Dated material, Lab No	Reference
Northern and northwestern Estonia					
Lake Tänavjärv (18.4)	316–321	4490±70	4920–5310 (5130±100)	Gyttja Tln–3306	Paper II
	321–324	4600±100	5050–5450 (5270±110)	Gyttja Tln–3305	Paper II
	342–347	4930±40	5600–5730 (5660±40)	Wood Poz–42173	Paper II
Lake Klooga (11.8)	335–340	3760±40	3990–4230 (4110±60)	Plant remains Poz–42168	Paper II
	390–395	3840±50	4150–4420 (4280±80)	Plant remains Poz–42169	Paper II
Lake Lohja (5.8)	365–370	2280±30	2160–2350 (2270±60)	Bark Poz–42171	Paper I
	395	2490±35	2440–2730 (2580±80)	Bark Poz–42172	Paper I
Lake Käsmu (3.9)	416	1830±30	1700–1860 (1770±40)	Bark Poz–42177	Paper I
	433	1910±30	1790–1930 (1860±30)	Wood Poz–42166	Paper I
Lake Harku (1.2 m)	275–280	1185±30	1065–1170 (1120±55)	Plant remains Poz–51453	Paper III
	295–300	1265±30	1175–1260 (1220±45)	Plant remains Poz–51454	Paper III
	424	1895±35	1750–1890 (1820±70)	Wood Poz–49185	Paper III
Island of Hiiumaa					
Kõivasoo Bog (27.5)	213–223	6580±60	7430–7580 (7510±50)	Gyttja Ta–527	Königsson et al. 1998
	245.5	6830±90	7750–8000 (7890±60)	Plant remains Ua–12071	Königsson et al. 1998
	245–255	7440±60	8020–8250 (8120±70)	Carbonate fraction Ta–528	Königsson et al. 1998

	260–270	7850±70	8410–8610 (8490±50)	Carbonate fraction Ta–529	Königsson et al. 1998
	315–325	8190±90	9170–9490 (9350±90)	Carbonate fraction Ta–530	Königsson et al. 1998
Loopsoo Bog (21.5–22)	365–366	5190±40	5430–5570 (5490±40)	Peat Poz–52916	Paper IV
	370–371	5050±40	5450–5570 (5520±30)	Peat Poz–52917	Paper IV
	377–382	5600±40	5890–6100 (5980±60)	Plant remains Poz–52918	Paper IV
	419	1240±30	1070–1270 (1180±60)	Wood (root?) Poz–50760	Paper IV
	450–455	5140±40	5750–5990 (5670±70)	Plant remains Poz–50763	Paper IV
Lake Tihu Keskmine (14.5)	186	4470±40	4900–5270 (5080±90)	Twig Poz–52920	Paper IV
	190–195	1120±40	940–1170 (1030±60)	Bark Poz–50761	Paper IV
	200–204	4490±40	5050–5300 (5200±70)	Plant remains Poz–52921	Paper IV
	210–215	2960±100	2870–3370 (3120±130)	Plant remains Poz–52922	Paper IV
Lake Prassi (7.2)	110	1020±30	830–1050 (940±30)	Plant remains Poz–52915	Paper IV
	114	1550±30	1310–1430 (1370±30)	Wood Poz–50758	Paper IV
	123	1480±40	1350–1510 (1410±50)	Bark Poz–50759	Paper IV
	139	1490±30	1360–1520 (1420±50)	Wood Poz–52914	Paper IV
	189	1630±30	1420–1600 (1520±50)	Wood Poz–52913	Paper IV
Southern coast of the Gulf of Riga					
Lake Lilaste (0.5)	380	2385±30	2680–2340 (2420)	Gyttja Poz–63856	Paper V
	515–520	3793±60	4410–3990 (4190)	Gyttja Tln–3500	Paper V
	540–545	3920±55	4520–4160 (4350)	Gyttja Tln–3497	Paper V
	660	4880±30	5710–5490 (5620)	Gyttja Poz–63857	Paper V

	878	6370±40	7420–7250 (7310)	Gyttja Poz–57601	Paper V
	1050	7810±35	8690–8480 (8590)	Gyttja Poz–63858	Paper V
	1222	9630±60	11190–10760 (10980)	Wood Poz–48438	Paper V
	1240	9710±50	11240–10810 (11110)	Wood Poz–48439	Paper V
Lake Ataru (3.2)	935–937	8150±50	9260–9000 (9090)	Twig Poz–56240	Current study
	1097– 1099	9570±50	11130–10720 (10930)	Twig Poz–56239	Current study
Lake Laveru (1.2)	672	5170±40	6640–6400 (6500)	Twig Poz–56229	Current study
	772	9220±50	10520–10240 (10380)	Wood Poz–56230	Current study
	548–552	5290±35	6190–5940 (6080)	Bark Poz–56231	Current study
Lake Pulkstenu (1.9)	738–740	9100±50	10400–10190 (10250)	Bark Poz–56233	Current study
	774	9640±50	11200–10780 (10970)	Wood/bark Poz–56235	Current study
Lake Jugu (4.1)	1070	4890±35	5720–5580 (5620)	Twig Poz–56232	Current study
Lake Linu (3.3)	745–747	5270±35	6180–5930 (6060)	Bark Poz–56236	Current study
	816	6810±40	7710–7580 (7640)	Bark Poz–56237	Current study
Lake Slokas (1.3)	223	7460±90	9420–8040 (8270)	Wood Poz–48440	Current study

GIS and software applied

The palaeogeographic maps are based on geographic information system (GIS) analysis in which the interpolated water level surfaces were removed from the digital terrain model (DTM; Rosentau et al. 2009). Topographical maps on scales of 1:10 000, 1:25 000, and 1:50 000 were used to create DTMs with grid sizes of 20 m × 20 m for lakes Lohja and Käsnu (Papers I, II) and 15 m × 15 m for lakes Tānavjärvi and Klooga (Paper II). In Paper III, the DTM with a 10 m × 10 m grid size was based on LIDAR (Estonian Land Survey). The peat deposits were removed from the DTM according to the soil maps of 1:10 000 scale (Estonian Land Survey) and unpublished reports from the Estonian Geological Survey. GIS-based water-level surfaces for BIL (11 700 cal BP), YS (11 650 cal BP), AL (10 300 cal BP), and LitS (7800 cal BP) were derived from the Estonian coastal formation database (Saarse et al. 2003, 2006). The water-level surfaces were created with ±1 m residual; thus, the reconstructed mean water level can fluctuate about 2 m. The LitS water level surface (7800 cal BP) was modified according to the assumption that the relative sea level has regressed evenly due to linear land

uplift (Mörner 1979; Lindén et al. 2006), and new water level surfaces for 8500, 7100, 6800, 5100, 4800, 4400, 2700 and 2200 cal BP were interpolated.

The RSL curves were compiled for the northern Estonia (Papers I, II), Hiiumaa (Paper IV), and Riga area (Paper V). The water level curves were based on modelled water levels of BIL (11 700 cal BP), YS (11 650 cal BP), AL (10 300 cal BP), and LitS (7800 cal BP) derived from the Estonian coastal formation database (Saarse et al. 2003, 2006) and complemented by lithological, biostratigraphical, and geochronological proxies.

LOI, MS, and diatom results were plotted by using Tilia v. 1.7.16 software (Grimm 2011), and all figures were combined and improved in CorelDRAW X4. In case of Lake Lilaste, the diatom zonation follows the constrained incremental sum of squares cluster analysis (Grimm 2011).

5. RESULTS AND DISCUSSION

5.1. Northern and northwestern Estonia

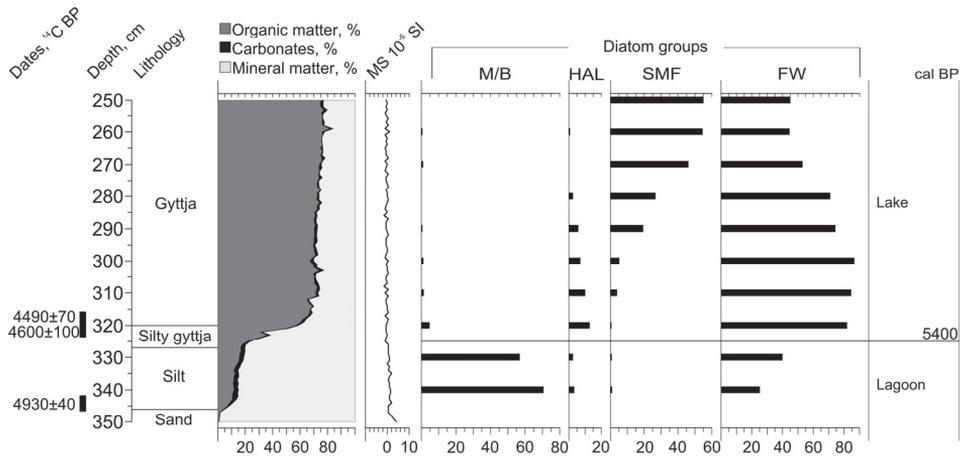
The studied lakes in north and northwestern Estonia are located within the Gulf of Finland drainage system on the terraces of the LitS and LimS at different altitudes between 18.4 m and 1.2 m a.s.l. (Table 1). During the transgression maximum of LitS at about 7800 cal BP, the coastal lowland was inundated by the sea, and only small islets emerged above the LitS waters. At that time, numerous beach ridge systems formed along the ancient shoreline, stretching continuously in the North Estonian Klint bays. The maximum heights of the shorelines of the LitS were recorded as 22.1 m a.s.l. at Tānavjārv, 21.9 m a.s.l. at Klooga, 22 m a.s.l. at Harku, 18.8 m a.s.l. at Lohja, and 17.7 m a.s.l. at Käsnu (Papers I, II, III). Owing to continuous land uplift, the coastal lowland gradually emerged from the sea and successively developed into the isolated freshwater lakes in the study area. GIS-based 3D palaeogeographic reconstructions and description of the local palaeoshoreline changes in northern and northwestern Estonia are presented in papers I–III.

The lithostratigraphy of the studied isolation basins are similar to some extent (Fig. 3). A common feature is the significant prevalence of MM content in sediments deposited during the sea or lagoonal phase. The beginning of the isolation, shown in the figures as the transition zone, can be distinguished after the increase in OM content and changes in diatom assemblages, which are excellent tools for determining the isolation contact. Furthermore, diatoms give additional information about environmental changes. During the isolation phase, newly emerged bare vegetation land was exposed to intensive erosion processes, thereby increasing the nutrient content in the water basin. Consequently, a fast sedimentation rate occurred in the studied isolation basins.

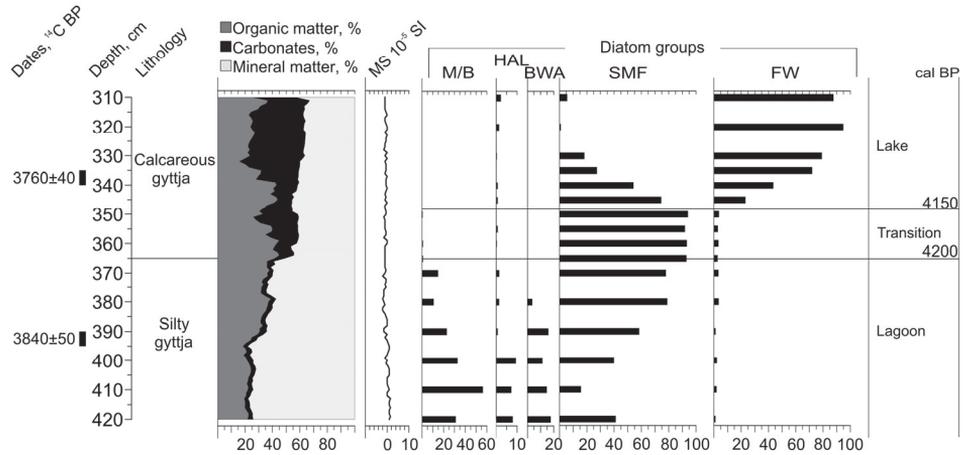
In subsequent paragraphs, the features of the isolation event in the studied lakes will be discussed. Detailed description and interpretation of the lithology and diatom assemblages of lakes Tānavjārv and Klooga are given in paper II; those for lakes Lohja and Käsnu are given in papers I, II; and those for Harku are given in paper III.

The isolation of **Lake Tānavjārv** was a rather short process. By 6500 cal BP, it was part of the shallow sea that formed about 500 years later into a lagoon with a passage in the south (Paper II). The distinct change in sediment lithostratigraphy from silty gyttja to gyttja and the shift in diatom composition from brackish to freshwater species in the sediment sequence of Lake Tānavjārv (Fig. 3A) are observed at 327 cm. This level is identified as the isolation contact. The age of the isolation is around 5400 cal BP. The palaeoreconstruction correlates with the conclusions drawn by Saarse et al. (1989) such that Lake Tānavjārv isolated from the LitS in the second half of the Atlantic period. A sharp decrease in marine/brackish diatom taxa suggests a rather short-term isolation from the sea. Initially, the water in the isolation basin was rich in nutrients, as confirmed by benthic diatoms such as *Navicula radiosa* and *Sellaphora pupula*. Subsequently,

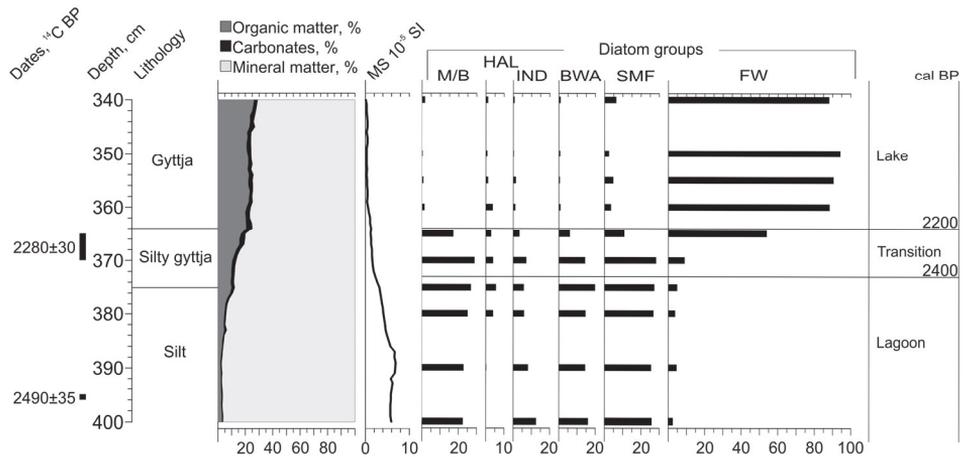
A Lake Tänävjärv



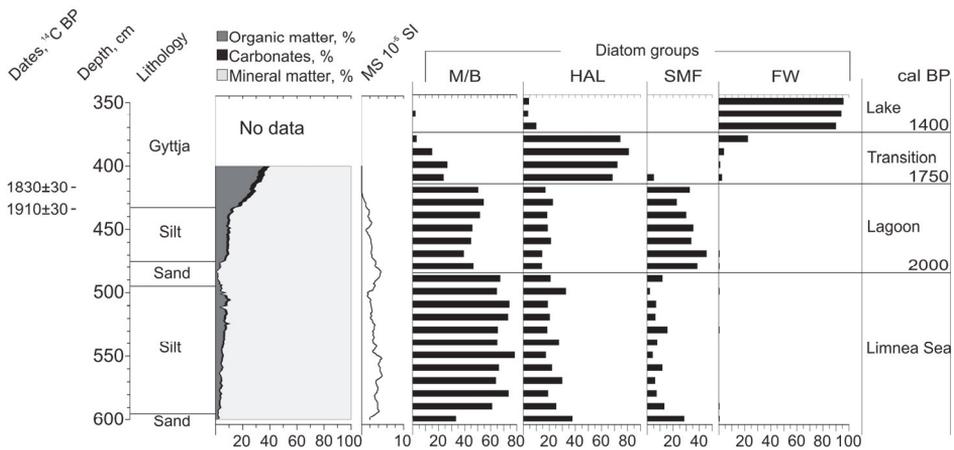
B Lake Klooga



C Lake Lohja



D Lake Käsü



E Lake Harku

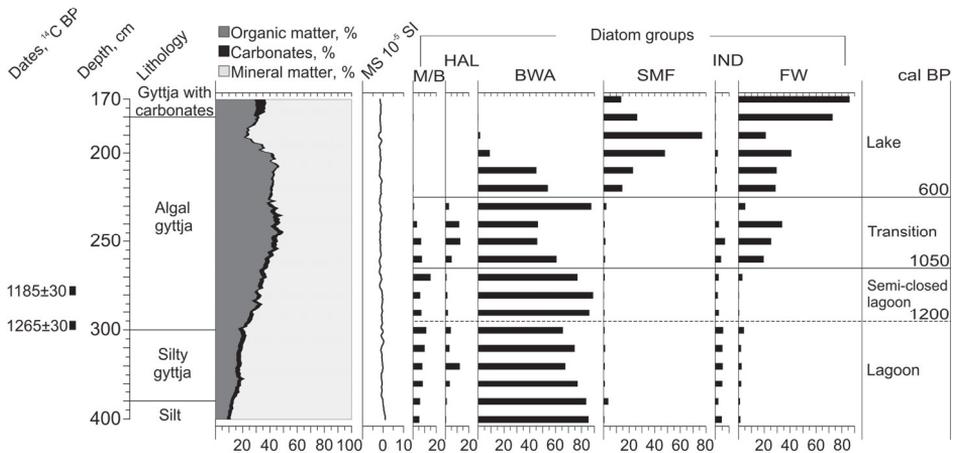


Figure 3. Sediment organic matter, carbonates and mineral matter content estimated by loss-on-ignition (% of dry weight), magnetic susceptibility (MS) and diatom summary diagram (% of total). **M/B** – marine/brackish taxa; **HAL** – halophilous taxa; **IND** – indifferent taxa; **BWA** – small-sized fragilarioid taxa with brackish water affinity; **SMF** – small-sized fragilarioid taxa; **FW** – freshwater taxa.

these diatom taxa were replaced by diatoms such as *Pinnularia* spp., *Brachysira serians*, *Eunotia* spp., and *Tabellaria fenestrata*, which indicate changes in the lake environment from alkaline to acidic, nutrient-poor conditions (Bigler et al. 2000; Dixit & Dickman 1986) owing to paludification of the lake catchment area.

Lake Klooga sediments contain OM rich silty gyttja that grades into calcareous gyttja (Fig. 3B). The carbonate content rises up to 33% during transition phase and up to 44% in the lake phase. The high carbonate content can be explained by limestone bedrock in the areas surrounding Klooga. The peak of small-sized fragilarioid taxa after the disappearance of the marine/brackish diatoms marks the transition to the final isolation of the basin. Thereupon, periphytic freshwater diatoms such as *Achnantheidium minutissimum*,

Cymbella falsata diluviana, and *Sellaphora vitabunda* indicate the establishment of a shallow, hard-water lake environment. The isolation event occurred around 4200 cal BP. The isolation of Lake Klooga was a rather slow process in comparison with that of Lake Tänävjärv. The isolation of Klooga, owing to the depth of the surrounding sea, was determined by the land uplift and by the development of various beach formations. Around 5000 cal BP, a beach ridge system developed to the east of the lake, whereas a tombolo began to form north of the lake. Although the diatom record confirms the isolation around 4200 cal BP, palaeoreconstructions indicate the closing of the northern connection with the sea 200 years later. This discrepancy could be explained by the higher water level in the lake, where the passage that connects the lake with the sea functioned as a drainage canal of the lake and prevented sea-water inflow into Klooga basin.

The lithostratigraphy of **Lake Lohja** follows the transition from silt to silty gyttja and to gyttja (Fig. 3C). The bottom layer of the silt is characterised by the absolute dominance of periphytic marine/brackish diatoms together with small-sized fragilarioid taxa that suggests a shallow sheltered lagoon-like basin connected with the Gulf of Finland through the open strait. A gradual increase of OM occurs between 375 cm and 365 cm and stabilises afterwards. This transition phase is characterised by the co-appearance of marine/brackish diatoms such as *Planothidium delicatulum*, *P. lemmermannii*, and *Navicula peregrina* and freshwater diatoms such as *Aulacoseira subarctica* and *A. granulata*. The marine/brackish diatom taxa replacement by planktonic freshwater diatoms indicates a rather abrupt isolation around 2200 cal BP.

The bottom part of the **Lake Käsü** sequence represents LimS sediments consisting of sand and silt overlain by sand with an OM rich layer and silt with OM deposited in the lagoonal phase (Fig. 3D). A typical feature of the changing environment is the mass occurrence of small-sized fragilarioid taxa (Stabell 1985). The Käsü *Fragilaria* spp. flourished during the lagoonal phase and disappeared just prior the isolation. The transition was indicated after the gradual increase of OM, decrease of marine/brackish diatoms, peak of planktonic halophilous *Cyclotella meneghiniana*, and appearance of planktonic freshwater *Aulacoseira granulata*. The diatom composition after the isolation, mainly *A. granulata*, *A. subarctica*, and *Cyclostephanos dubius*, reflects the development of a small eutrophic lake. According to AMS ¹⁴C ages, the final isolation of Käsü occurred around 1800 cal BP. Although Kessel et al. (1986) stated that Lake Käsü was isolated around 800 cal BP, new evidence confirms that the isolation occurred about 1000 years earlier.

In summary, lakes Lohja and Käsü maintained a broad connection with the sea until 4000 cal BP, and semi-closed lagoons with narrow passages between the sea and the lagoons formed by 3000 cal BP. The isolation of Lohja basin was determined mainly by the post-glacial relief-forming processes, primarily by beach ridges. However, the isolation of Lake Käsü was controlled by glacial formations, primarily by a buried esker ridge to the west of the lake.

In the sediment sequence of **Lake Harku**, the amount of OM increases gradually, reaching its maximum at 255–230 cm and covering the time span of

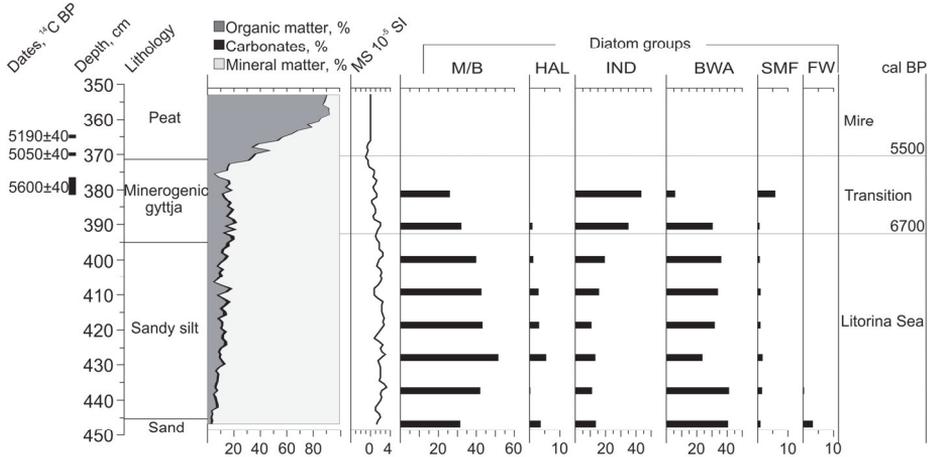
980–630 cal BP (Fig. 3E). The increase in OM indicates high primary production in the water basin. In addition, the increase in planktonic halophilous *Cyclotella meneghiniana* and planktonic freshwater *Stephanodiscus parvus* confirms the establishment of a nutrient-rich environment (Anderson 1990; Weckström & Juggins 2006; Witak 2013). Several factors and their interactions can explain the enhanced nutrient content in Lake Harku such as the occasional mixing of brackish and fresh water, which promotes biological productivity and enrichment in organic compounds (Head 1976); an intense nutrient input from the sparsely vegetated catchment area (Seppä et al. 2000); and climate amelioration during the Medieval Climate Optimum (Hughes & Diaz 1994; Mann et al. 2009). The same sediment layer represents the transition phase prior to the isolation that, according to the palaeogeographic model, occurred around 800 cal BP. This age is in disagreement with the diatom assemblages, indicating that brackish-water conditions lasted longer, at least over the next 300 years. Several factors could have contributed to the occasional seawater intrusions, such as the plain landscape, low-lying threshold, and proximity to the sea. Previous studies by Saarse (1994) reported that the Harku basin isolated from the sea considerably earlier, around 1500 years ago. However, this estimation was incorrect because it was based on only pollen data, lithostratigraphy, and morphology of coastal formations; no radiocarbon dates were used.

5.2. Island of Hiiumaa

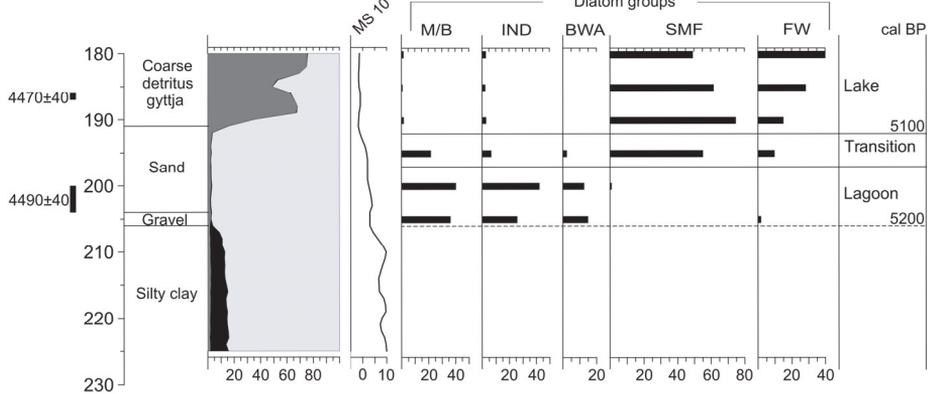
Emerging from the sea after the BIL drainage about 11 000 cal BP, the Island of Hiiumaa is rather young in its present shape (Paper IV). As a result of the rapid land uplift, the AL and LitS left behind numerous ancient coastal formations such as beach ridges, a wide spectrum of scarps, ancient lagoons, and dune systems. Therefore, it is one of the best regions for studying water-level changes. Three isolation basins, Loopsoo Bog and lakes Tihu and Prassi at different elevations of 21.5 m, 14.5 m, and 7.2 m a.s.l., respectively, were chosen as potential sites to study and reconstruct the shoreline displacement curve of the Island of Hiiumaa. In order to provide more precise reconstructions and to cover the time period from the beginning of the emergence of Hiiumaa, data from the earlier studied Kõivasoo Bog, with a threshold at 27.5 m a.s.l. (Königsson et al. 1998; Saarse et al. 2000), were incorporated in this study. Detailed descriptions and interpretations of the lithology and diatom assemblages of Kõivasoo, Loopsoo, Lake Tihu, and Lake Prassi, as well as GIS-based 3D palaeogeographic reconstructions and description of the local palaeoshoreline changes in the Island of Hiiumaa, are given in Paper IV.

The **Kõivasoo** basin isolated twice (Paper IV). The first isolation occurred prior to 9400 cal BP to form a shallow, hard-water coastal lake. A shift from periphytic shallow-lake taxa to planktonic large-lake taxa such as *Aulacoseira islandica* and *Stephanodiscus neoastraea* indicates the water level rise and the establishment of the new connection between Kõivasoo basin and the Baltic Sea basin around 9350 cal BP. Hence, during the Early LitS stage, a freshwater lagoon was formed. The second isolation from the LitS occurred around 8550 cal BP

A Loopsoo Bog



B Lake Tihu Keskmine



C Lake Prassi

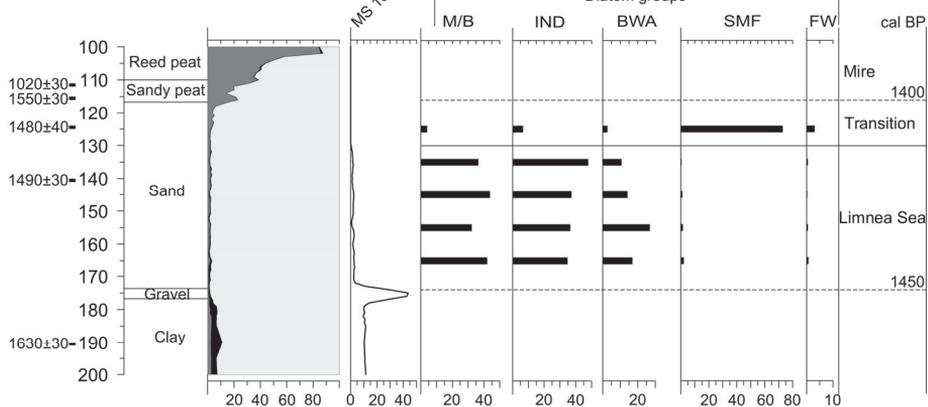


Figure 4. Sediment organic matter, carbonates and mineral matter content estimated by loss-on-ignition (% of dry weight), magnetic susceptibility (MS) and diatom summary diagram (% of total). **M/B** – marine/brackish taxa; **HAL** – halophilous taxa; **IND** – indifferent taxa; **BWA** – small-sized fragilarioid taxa with brackish water affinity; **SMF** – small-sized fragilarioid taxa; **FW** – freshwater taxa.

(Saarse et al. 2000). This final isolation is indicated by the disappearance of large-lake diatoms, a decrease in littoral *Mastogloia* spp., and replacement by small, shallow-lake taxa.

According to the diatom data and AMS ^{14}C dates, the sandy silt in the **Loopsoo** basin was deposited in the LitS (Fig. 4A). Periphytic marine/brackish diatoms such as *Hyalodiscus scoticus*, *Cocconeis scutellum*, and *Planothidium delicatulum* indicate a shallow depositional environment during the LitS stage. The isolation contact is not sharp in the Loopsoo sequence. Marine/brackish and small-sized fragilarioid taxa with brackish-water affinity were gradually replaced by indifferent taxa. Moreover, the deposition of minerogenic gyttja suggests that isolation was an extended process that started around 6800 cal BP and terminated around 6000 cal BP. During the transitional phase, Loopsoo basin was strongly affected by the sea because it was located on a small island in the open sea (Paper IV). Because the upper part of the minerogenic gyttja and peat were barren of diatoms, the final isolation might have occurred prior to the onset of peat accumulation at 5500 cal BP. However, such long-lasting isolation is in conflict with the water-level simulation results, which show that isolation likely occurred around 6500 cal BP. Such a discrepancy could be explained by two factors: 1) proximity of the Loopsoo basin to the sea and 2) the location on a small island exposed to wind and wave activity (Paper IV). These storm surges and high wave activity caused the sea water to flood the basin during and after the isolation, which complicates the determination of the exact isolation contact.

During the LitS stage, typical epiphytic and epipsammic marine diatoms such as *Catenula adhaerens*, *Diploneis didyma*, and *Actinoptychus octonarius* inhabited the shallow overgrowing **Tihu** lagoon. The transition zone from a lagoon to a freshwater basin (Fig. 4B) is marked by a peak of small-sized fragilarioid taxa, indicating a change in the depositional environment (Seppä & Tikkanen 1998; Seppä et al. 2000; Risberg et al. 2005). The replacement of sand by coarse detritus gyttja together with the increase in freshwater diatoms at Tihu mark the isolation contact around 5100 cal BP.

Diatoms in the **Lake Prassi** sequence are preserved only in the middle part of the sand layer, whereas the lower and upper parts of sand are barren of diatoms (Fig. 4C). The peak of small-sized *Fragilaria* spp. indicates the beginning of the transition from the LimS to a somewhat isolated water body. The absence of diatoms in the upper part of the sequence complicates the determination of the exact isolation contact. If the final isolation contact is assumed to be the border between sand and sandy peat, according to ^{14}C dates, Prassi was isolated at about 1400 cal BP. However, this age is in conflict with the water-level simulation results, which revealed isolation more than 1000 years earlier, at ca 2500–2700 cal BP. Therefore, it should be considered that the dated wood might have been redeposited or contaminated during the coring process. Furthermore, the uppermost part of the sand with abundant woody pieces could have been carried into the basin during storms, when the water level may have risen up to 3 m (Paper IV).

5.3. Southern coast of Gulf of Riga

A complex interplay between various processes such as glacio–isostasy, eustatic sea-level rise, those of large river systems (e.g. river incision, river bifurcation, deposition of suspended sediments), longshore sediment transport, and aeolian processes formed the coastal area of the Gulf of Riga. In this thesis, the focus is on eastern Rigavas Plain, the vicinity of the mouth of the River Gauja, and the area between Gauja and Daugava rivers. Eight lakes were chosen to reconstruct the water level changes in the southern part of the Gulf of Riga (Fig. 5). In order to facilitate the interpretation of obtained data, these lakes were grouped into the following three categories according to similarities in the sediment sequences:

- 1) lakes Lilaste and Dunu, with the longest sediment sequences consisting mostly of various types of gyttja (possibly LitS intrusions);
- 2) lakes Laveru, Linu, and Jugu, which include sediment sequences with a considerable layer of sandy silt with mud and dispersed OM (possibly river intrusions); and
- 3) lakes Ataru, Pulkstenu, and Slokas, which include sediment sequences with peat layers at the bottom (possibly dry climatic periods).

The sediment sequences of **lakes Lilaste** and **Dunu** contain mostly gyttja, including calcareous and silty gyttja (Fig. 5). Although both lakes are connected to each other and have rather similar lithostratigraphy, Lake Lilaste was chosen for detailed study because it is located closer to the coast and hence has a greater possibility of having a stronger signal of marine water inflows.

Clayey silt at the basal part of the sequence of Lilaste (Fig. 6), most likely deposited in BIL, and sand might have been deposited during the drainage of BIL or erosion at the time of the YS. Two AMS ¹⁴C-dated wood remains from layers of sand (at 1240 cm) and thin gyttja (at 1222 cm) of lake Lilaste at $11\,160 \pm 60$ cal BP and $10\,890 \pm 80$ cal BP (Table 4), respectively, indicate the possibility that these sediments accumulated during the YS phase. The wood remains suggest that a low water level existed in the area during the YS phase. Numerous plant macroremains and diatoms in the gyttja layer (at 1218–1226 cm) indicate a shallow eutrophic freshwater lake environment and suggest that the Lilaste basin was not connected with the Baltic Sea basin at the end of YS phase. Most likely, the AL transgression at about 10 300 cal BP (Saarse et al. 2003) induced a groundwater-table increase in coastal areas, which successively caused an increase in lake-water depth. As the water level gradually began to rise, carbonate-rich till situated close to the eastern shore caused lacustrine lime precipitation that was replaced by an accumulation of silty gyttja at ca 9750 cal BP. The lithostratigraphy shows distinct OM and MM fluctuations from 8700 cal BP to 4200 cal BP (Fig. 6), indicating changes in the sedimentation environment. Diatom evidence for that time interval shows a dominance of freshwater diatoms over marine/brackish diatoms (Fig. 6), which suggests that Lake Lilaste during the LitS stage was a freshwater lake with intermittent marine-water intrusions. Two remarkable episodes of intense marine-water inflow, confirmed by a sharp increase in diatom assemblages preferring brackish water or environment with

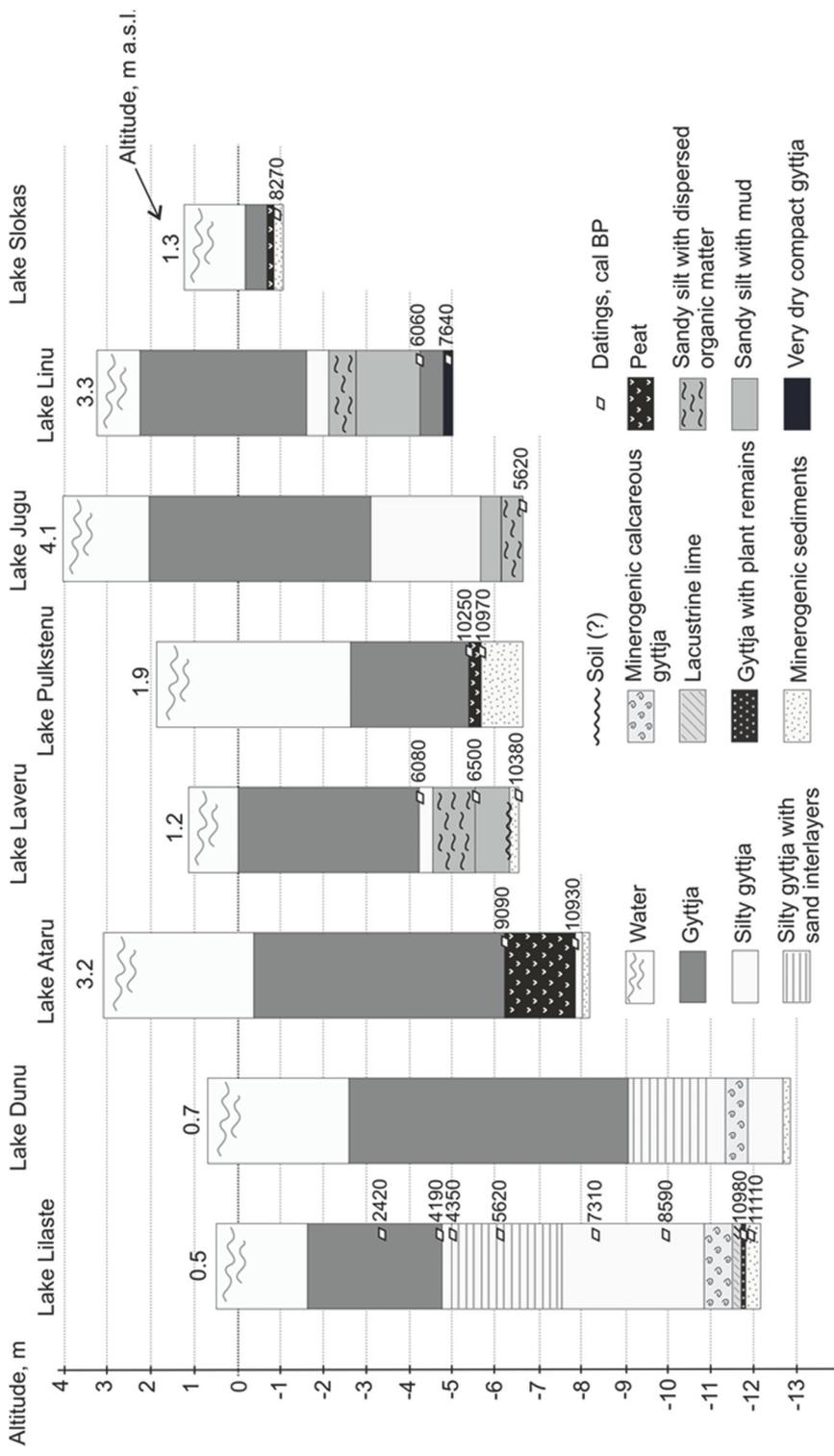


Figure 5. Simplified lithologic description of studied sites in the southern coast of Gulf of Riga.

A Lake Lilaste

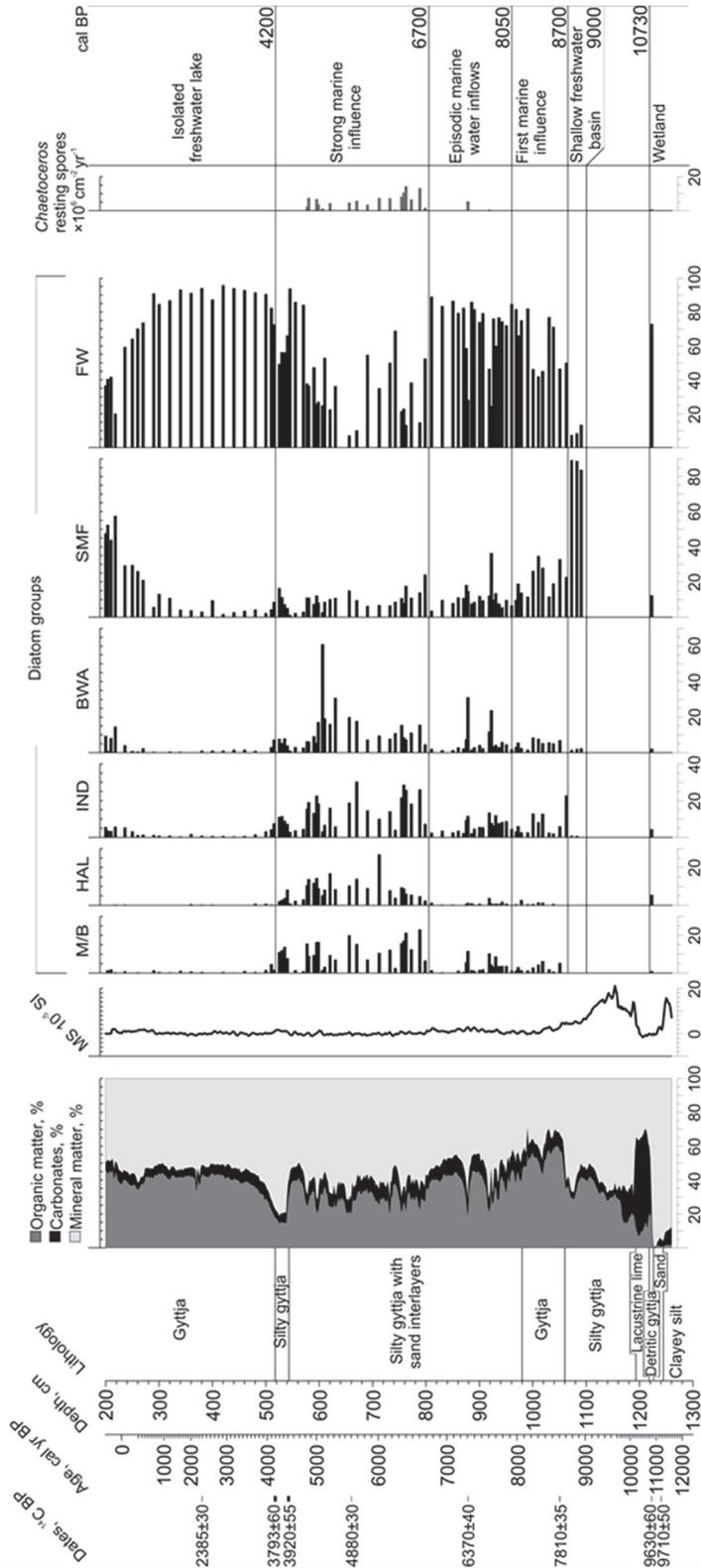


Figure 6. Sediment organic matter, carbonates and mineral matter content estimated by loss-on-ignition (% of dry weight), magnetic susceptibility (MS) and diatom summary diagram (% of total). **M/B** - marine/brackish taxa; **HAL** - halophilous taxa; **IND** - indifferent taxa; **BWA** - small-sized fragiliaroid taxa with brackish water affinity; **SMF** - small-sized fragiliaroid taxa; **FW** - freshwater taxa.

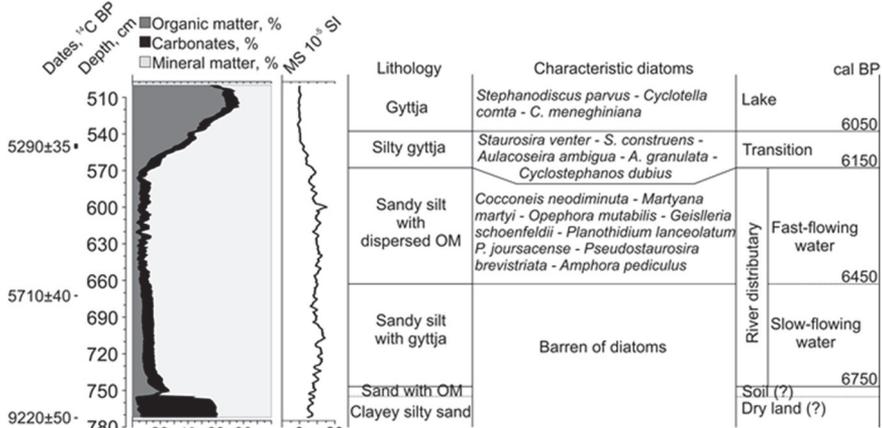
high-conductivity, occurred at 7600 and 7300 cal BP. Furthermore, significant increases in marine/brackish (*Cyclotella choctawhatcheeana*, *Achnanthes fogedii*, *Navicula gregaria*, *N. perminuta*, *Planothidium delicatulum*, *P. lemmermannii*), halophilous (*Cyclotella meneghiniana*, *Hippodonta hungarica*, *Navicula clementis*), indifferent (*Amphora pediculus*, *Cocconeis neothumensis*, *Diatoma tenuis*, *Epithemia sorex*, *Rhoicosphenia abbreviata*) and small-sized fragilarioid taxa with brackish-water affinity (*Fragilaria sopotensis*, *Martyana atomus*, *Staurosira punctiformis*), as well as an appearance of *Chaetoceros* resting spores from 6700 up to 4200, indicates intense marine-water intrusion. A comprehensive discussion about the causes of the notable marine-water inflows are discussed in paper V. A gradual increase in OM occurred after the final isolation around 4200 cal BP, confirming the formation of a stable freshwater environment in the lake. In addition to lithological evidence, the disappearance of marine/brackish diatoms and the increase in planktonic freshwater *Aulacoseira* spp. such as *A. ambigua*, *A. granulata*, *A. islandica*, and *A. subarctica* confirm the final isolation at 4200 cal BP (Paper V).

In summary, the lithology and diatom data of the Lake Lilaste sequence sustain the hypothesis regarding long-lasting, intermittent sea-water inflow for approximately 4500 years, covering the time span of 8700–4200 cal BP.

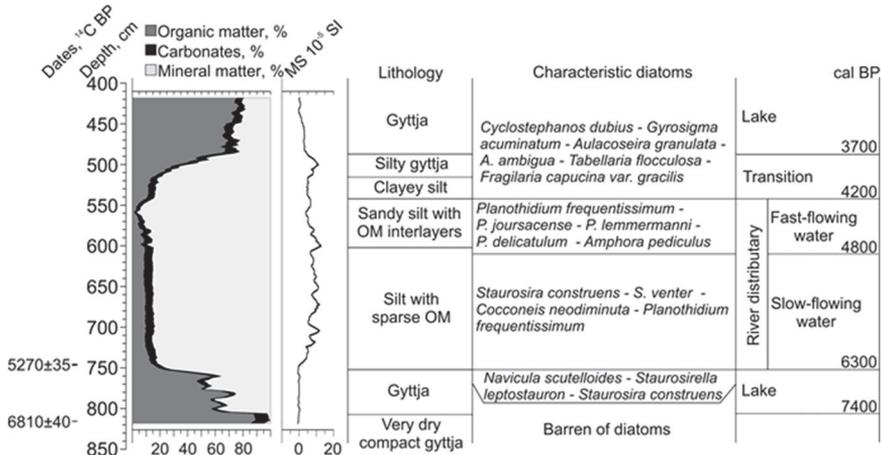
During fieldwork in lakes **Laveru**, **Jugu**, and **Linu**, peculiar sediment layers were detected such as sandy silt with mud and sandy silt with dispersed OM (Fig. 5). The sediment layers are visually similar, and the lakes are located near each other in the area between the Gauja and Daugava rivers. Therefore, the same origin is assumed, and the silty sand was deposited under the same or almost analogous environmental conditions. To determine the sedimentation environment, multi-proxy analyses such as LOI, MS, diatom, and grain size were applied. The lithology description, LOI, and MS are presented in Fig. 7, and the dominant diatom species are shown in Fig. 8.

A piece of wood dated to 10 380 cal BP from the minerogenic sediments at the bottom of the sediment sequence of **Lake Laveru** at a core depth of 772 cm (Fig. 7A) supports the idea of low sea level during the AL stage. The distinct boundary between minerogenic sediments and sand with OM at a core depth of 755 cm, which might be remnants of ancient soil, is considered as a hiatus. Unfortunately, an insufficient amount of radiocarbon dates prevents the precise determination of the soil layer age. Sand with OM, considered as ancient soil, might have formed during AL and early LitS. Above the boundary, sandy silt with mud is embedded. The results of LOI and grain size analysis imply a rather stable sedimentary environment which could be a distributary of the Gauja River delta. In that area, slow-flowing water with abundant suspended material promoted rapid sediment accumulation. Unfortunately, diatoms are absent from this layer. Fluctuations in MM content and the increase of the amount of sand suggest a transition to a more dynamic environment at around 6500 cal BP. Laveru basin as the marginal part of the distributary existed until ca 6100 cal BP.

A Lake Laveru



B Lake Linu



C Lake Jugu

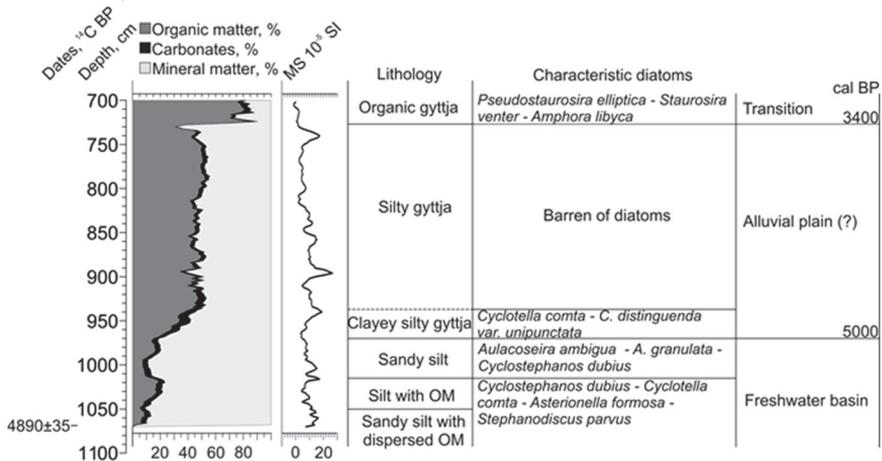
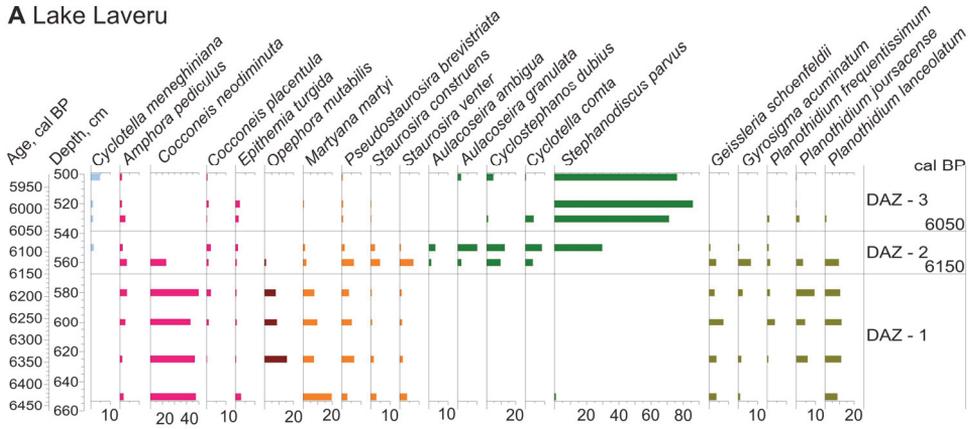
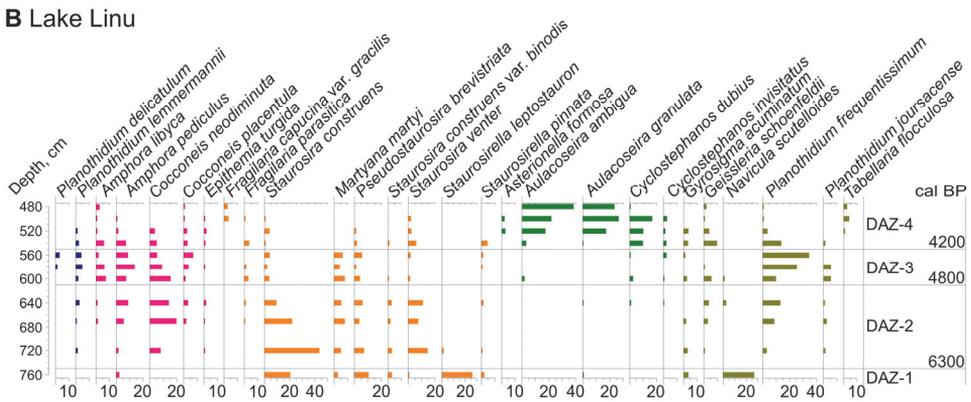


Figure 7. Sediment organic matter, carbonates and mineral matter content estimated by loss-on-ignition (% of dry weight), magnetic susceptibility (MS) and characteristic diatoms.

A Lake Laveru



B Lake Linu



C Lake Jugu

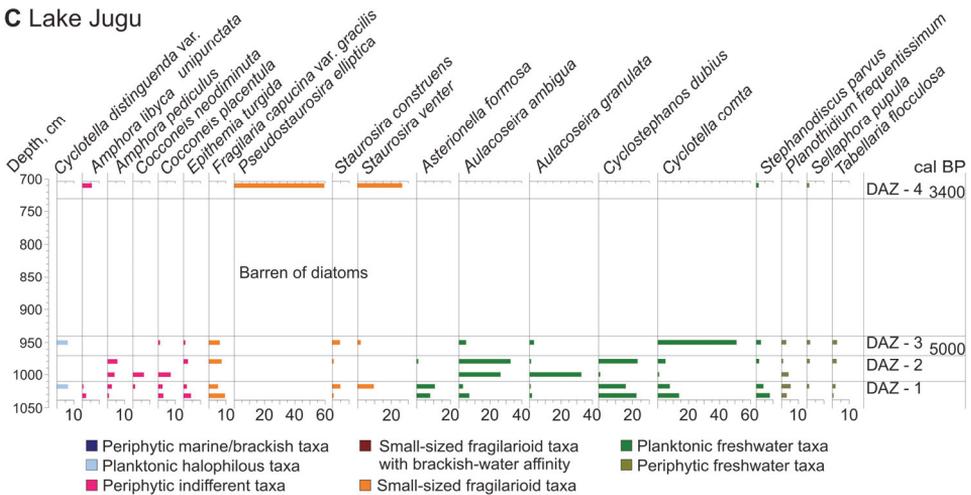


Figure 8. Percentage diagrams of selected diatom taxa from **A** – Lake Laveru, **B** – Lake Linu and **C** – Lake Jugu. DAZ - diatom assemblage zone.

The beginning of the isolation is marked by a shift in diatom assemblages and thus periphytic indifferent taxa replacement by planktonic freshwater diatoms such as *Aulacoseira granulata*, *Cyclostephanos dubius*, *Cyclotella comta*, and *Stephanodiscus parvus* (Fig. 8A) in addition to a gradual increase in OM from 10% to 40%. The establishment of a eutrophic freshwater lake, or the final isolation, occurred around 6000 cal BP.

Very dry compact gyttja with a high OM content of 85–90% is considered to be evidence of a freshwater basin during the LitS until 6300 cal BP, when the river water might have flooded the **Linu** basin (Fig. 7B). The change from standing water to a flowing water environment is confirmed by the rapid decrease in OM content and changes in diatom assemblages. The most frequent diatom taxa found in the lower part of the silt with sparse OM are small-sized *Fragilaria* spp. (Fig. 8B), such as *Staurosira construens* and *S. venter*, which were gradually replaced by periphytic indifferent *Cocconeis neodiminuta* and *Amphora pediculus* and periphytic freshwater *Planothidium frequentissimum*. The appearance of sand in silty deposits at 602 cm, as well as the dominance of *Planothidium* spp., which often are found in littoral zones attached to sand grains, and periphytic *Amphora pediculus*, *A. libyca*, *Cocconeis neodiminuta*, and *C. placentula* suggest alteration to a more dynamic water environment. The identified diatom species together with changes in lithology suggest a slow-flowing water environment transition to a more dynamic environment at about 4800 cal BP, shown in Fig. 8 as a river distributary. According to the lithology and diatom data, isolation of Linu basin occurred at 4200–3700 cal BP. Therefore, it can be concluded that a delta gradually filled by sediments formed at the end of the LitS stage. Therefore, the size of the river delta affected by river water decreased, and the shoreline moved towards the sea. Alternatively, long-shore sediment transport carried additional material to form a wider beach area.

The basal minerogenic sediments were not reached in the **Lake Jugu** core owing to the density and deep coring depth of 1070 cm. The lithostratigraphy of the sequence of Lake Jugu (Fig. 7C) differs from that in lakes Laveru and Linu. The OM content is higher than that in the two other lake sequences. The diatom species of *Cyclostephanos dubius*, *Cyclotella comta*, *C. distinguenda* var. *unipunctata*, *Aulacoseira ambigua*, and *A. granulata* in the silt with OM and sandy silt indicate a freshwater environment with a high nutrient content and conductivity (Fig. 8C). Around 5000 cal BP, the OM content increased, and planktonic freshwater *Cyclotella comta* and *C. distinguenda* var. *unipunctata* prevailed. A layer of silty gyttja 220 cm in thickness that was deposited until 3400 cal BP is barren of diatoms. Fluctuations in MM suggest that Jugu basin was part of a regularly flooded alluvial plain in which the conditions were not appropriate for diatom preservation. The transition phase from alluvial plain to isolated freshwater lake environment is indicated by the rapid increase in OM from approximately 50% to 80% and the mass occurrence of small-sized fragilarioid taxa such as *Pseudostaurosira elliptica* and *Staurosira venter*. The location of Lake Jugu, along with LOI and diatom data suggest that its basin might be located on the river island formed through bifurcation.

Although the diatoms identified in the sequences of lakes Laveru and Linu do not clearly show the environment that existed during the deposition time, many of these species often are found in flowing waters. For example, *Amphora pediculus*, *Cocconeis placentula*, *Cyclotella meneghiniana*, and *Planothidium delicatulum* are frequently found in many rivers; *Cocconeis neodiminuta* are commonly observed in freshwater lakes, and ponds and in slow-flowing waters; and *Planothidium frequentissimum*, *P. lanceolatum*, *Staurosira construens* var. *binodis*, *S. venter*, and *Epithemia turgida* usually inhabit standing waters but can also be found in rivers (Krammer & Lange-Bertalot 1988, 1991a, 1991b; Witkowski 1994). Identified diatom assemblages such as *Amphora libyca*, *A. pediculus*, *Cyclotella meneghiniana*, *C. distinguenda* var. *unipunctata*, *C. comta*, *Epithemia turgida*, *Gyrosigma acuminatum*, *Planothidium frequentissimum*, *P. joursacense*, and *Geissleria schoenfeldii* represent a freshwater environment with moderate or high electrolyte concentration, and particular diatom species such as *Planothidium delicatulum*, *P. lemmermannii*, *Opephora mutabilis*, *Pseudostaurosira brevistriata*, *Staurosirella pinnata*, and *Cyclostephanos dubius* suggest possible intermittent brackish water influence (Krammer & Lange-Bertalot 1986, 1988, 1991a, 1991b; Witkowski 1994).

In summary, the sea-level rise caused river bifurcation; hence, Laveru and Linu basins became parts of the distributaries of the Gauja River delta during the LitS. Owing to its altitude, Lake Jugu might have been located on a river island, which was regularly inundated by spring floods. Because marine water entering the river mouth is presently observed during low river discharge and strong wind gusts (Eberhards 2003), it is assumed that intermittent LitS brackish water inflows promoted the presence of diatoms that tolerate moderate to high electrolyte contents. Consequently, LitS did not reach the threshold of Laveru, Linu, and Jugu basins.

A common feature of the sediment sequences of lakes **Ataru**, **Pulkstenu**, and **Slokas** is the lowermost peat layer lying just above basal minerogenic sediments (Fig. 5). The peat was interpreted as *in situ* accumulated deposits of terrestrial origin. Therefore, these layers are good indicators of water level changes because the formation of basal peat in coastal areas depends on groundwater level directly controlled by the altitude of the sea. According to the AMS ¹⁴C ages, the peat accumulated in Lake Pulkstenu at 11 000–10 300 cal BP and in Lake Ataru at 10 900–9100 cal BP, which suggests that sea level in the southern part of the Gulf of Riga during the YS, AL, and Early LitS stages did not exceed more than 6 m b.s.l. A similar peat layer at the bottom of the lake sediment sequence just above the basal minerogenic sediments in Lake Babelitis from the same study area (Fig. 2D) was reported by Danilans (1963). Although this peat layer was not dated, he assumed that it could have formed during the AL stage; hence, he concluded the water level of AL in the southern part of the Gulf of Riga could not be higher than 5–6 m b.s.l. The interruption of peat accumulation and deposition of gyttja in the coastal area of Rigavas Plain indicate a sea level rise. The sample of wood dated at 8270 cal BP (Table 4) from the basal minerogenic sediments of Lake Slokas indicates that the LitS water level at that time did not exceed 1 m b.s.l.

The peat layer just above the minerogenic sediments is embedded by gyttja containing *Campylodiscus clypeus*, which indicates LitS transgression. Unfortunately, there is a hiatus between the gyttja with brackish-water diatoms and lake sediments with freshwater diatoms; hence, it is not possible to determine the exact isolation contact in Lake Slokas.

The main conclusion of the above paragraphs is that the studied sites in the southern coast of Gulf of Riga do not confirm that the LitS level was higher than the modern sea level. The changes in sea level in this area over the Holocene are further discussed in the following subsection.

5.4. Sea-level reconstructions

In this chapter, changes of RSL in the eastern part of Baltic Sea and their causes are discussed. According to obtained data in combination with additional data from the Holocene Baltic Sea shoreline database (Saarse et al. 2003), RSL curves were compiled for northern and northwestern Estonia (Fig. 9), the Island of Hiiumaa (Fig. 10) and the southern coast of the Gulf of Riga (Fig. 11). The reconstructed water level curves in northern Estonia and the Island of Hiiumaa represent changes in RSL starting from the onset of AL to the present. Emphasis is placed on LitS and LimS because the study sites represent isolation from these particular stages.

North and northwestern Estonia

The reconstructed water-level curve is relatively regular from the onset of the LitS regression and differs considerably from the previous curves (Kents 1939; Kessel & Raukas 1979). The older RSL curves show up to five transgressional waves during the LitS and LimS stages; however, these fluctuations were not radiocarbon dated. The reconstructed water-level curve displays one LitS transgression at 7800 cal BP in northwestern Estonia (Fig. 9). The amplitude of the transgression is about 3–4 m, which is reconcilable with studies conducted in areas with a similar isobase of land uplift, such as southern Finland (Miettinen 2002; Miettinen & Hyvärinen 1997). A notably higher magnitude of LitS transgression occurred in areas of low land uplift of about 7 m in the area surrounding Pärnu (Veski et al. 2005) and about 5 m in the border area of Estonia and Latvia (Saarse et al. 2006). The isolation horizon provides regular upwards-younger radiocarbon ages, and the biostratigraphic proxies do not show evidence of transgressions after 6000 cal BP. The presented RSL curves show smooth decreases in sea level from 22 m a.s.l. down to the present level. Nevertheless, these curves do not rule out changes in local water level during storm surges.

The apparent uplift currently occurring in the Tänävjärv-Klooga and Harku areas is about 2.2 mm yr⁻¹; that in the Käsmu-Lohja area is 2.0 mm yr⁻¹ (Torim 2004). According to calculations, however, the land uplift rates were higher in the mid-Holocene, at 2.8 mm yr⁻¹ in the Tänävjärv-Klooga area and 2.5 mm yr⁻¹ and 2.4 mm yr⁻¹, in the Harku and Käsmu-Lohja areas, respectively, showing a continuous decreasing trend towards the present. These data are in harmony with

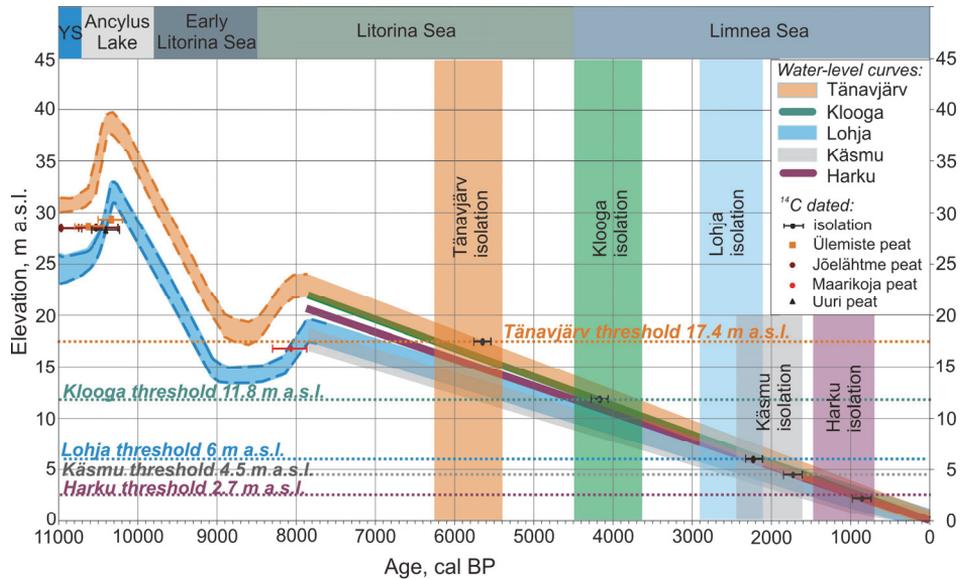


Figure 9. Water-level curves for the northern and northwestern Estonia, based on different proxies. Water levels for lakes Tānavjārv, Lohja and Kāsmu are shown with ± 1 m error bars, for lakes Klooga and Harku only mean modelled water level is shown.

the current rate of land uplift (Torim 2004), which also shows a decreasing trend.

In general, the reconstructed RSL curves show a linear trend of sea-level decrease that is in agreement with the curves recently reconstructed around the Gulf of Finland (Hyvärinen 1982; Seppä et al. 2000; Miettinen 2002, 2004; Rosentau et al. 2011; Saarnisto 2012) and the Gulf of Bothnia (Lindén et al. 2006; Widerlund & Andersson 2011). Changes in sea level over the last 5000 years have been reported from various regions of the world (Morhange et al. 2001; Gehrels et al. 2006; Murray-Wallace 2007), including the Baltic coast (Lampe 2005). If a rapid rise in the ocean level does not occur, the decreasing trend of the sea level and apparent uplift in northern Estonia will continue.

Island of Hiiumaa

The reconstructed water-level curves from the Island of Hiiumaa (Fig. 10) show a linear trend of sea-level decrease after the LitS transgression similar to that indicated by the reconstructed water-level curves from northern Estonia (Fig. 9). This trend also differs from the RSL curve presented earlier by Raukas & Ratas (1996) that proposed several LitS and LimS transgressions. Again, it should be kept in mind that gradual RSL lowering does not rule out minor changes within the error limits of the simulation. If large-scale water-level fluctuations would have occurred after the LitS maximum around 8800–8200 cal BP, they should have been visible in the diatom stratigraphy and lithostratigraphy of the studied sections. Data from the Island of Hiiumaa confirm that the RSL decreased consistently the last 8000 years owing to a progressively declining isostatic land uplift.

The reconstructed water-level curves display a time-transgressive LitS highstand peak occurring earlier in the Kõivasoo area and later in the Prassi area as a result of differences in the rate of land uplift. The calculated land uplift rate at 10 300 cal BP was 4.4 mm yr⁻¹ at Kõivasoo and decreased to 2.5 mm yr⁻¹ in recent time (Torim 2004). These results complement observations in northern Estonia (Paper I, II) as well as other studies from the Estonian coastal area (Veski et al. 2005; Saarse et al. 2010; Rosentau et al. 2013).

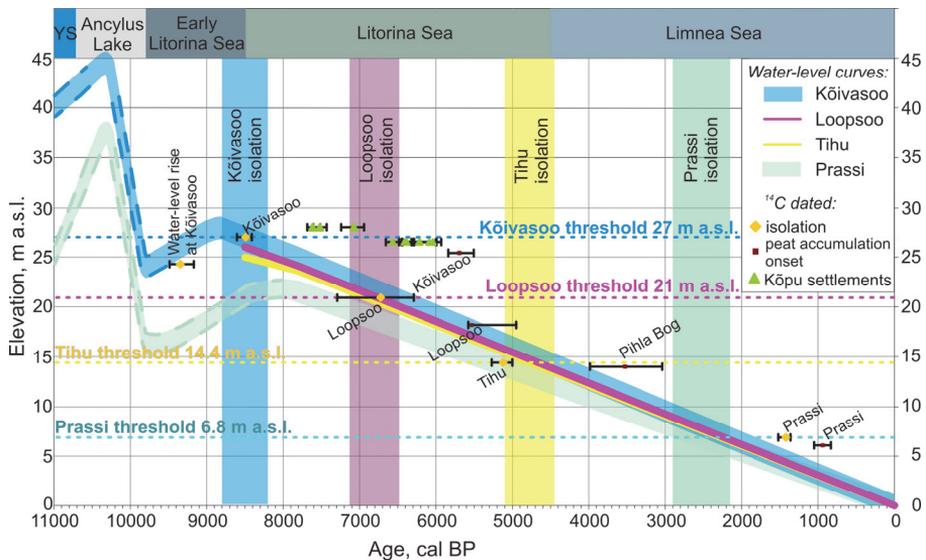


Figure 10. Water-level curves for the Island of Hiiumaa, based on different proxies. Water levels for Kõivasoo and Prassi are shown with ± 1 m error bars, for Loopsoo and Tihu mean modelled water level is shown.

Southern coast of the Gulf of Riga

The reconstructed water-level curves of the southern coast of the Gulf of Riga show changes in RSL (Fig. 11) over the Holocene. The surrounding areas of Riga were deglaciated and flooded by BIL at ca 14 000 cal BP (Vassiljev & Saarse 2013). Modelling results (Vassiljev & Saarse 2013) suggest that the water level reached up to 14–16 m a.s.l. just prior to the Billingen drainage at 11 700 cal BP (Figs. 11, 12). Considering that the water level after the rapid drainage of BIL was lowered by approximately 25 m to the ocean level, the YS maximum water level in the Gulf of Riga was approximately 9–11 m b.s.l. The land uplift in the Riga area most likely exceeded the ESL rise, resulting in YS regression. In earlier studies, Berzin (1967) and Kovalenko and Juskevics (1987) suggested that the YS minimum level was at least 29 m b.s.l. and 15 m b.s.l., respectively. Modelling results (Saarse et al. 2003) show that the water level in the Gulf of Riga during the AL transgression at 10 300 cal BP was about 12–17 m b.s.l.; thus, it can be concluded that the YS regression minimum was lower than that value (Fig. 11).

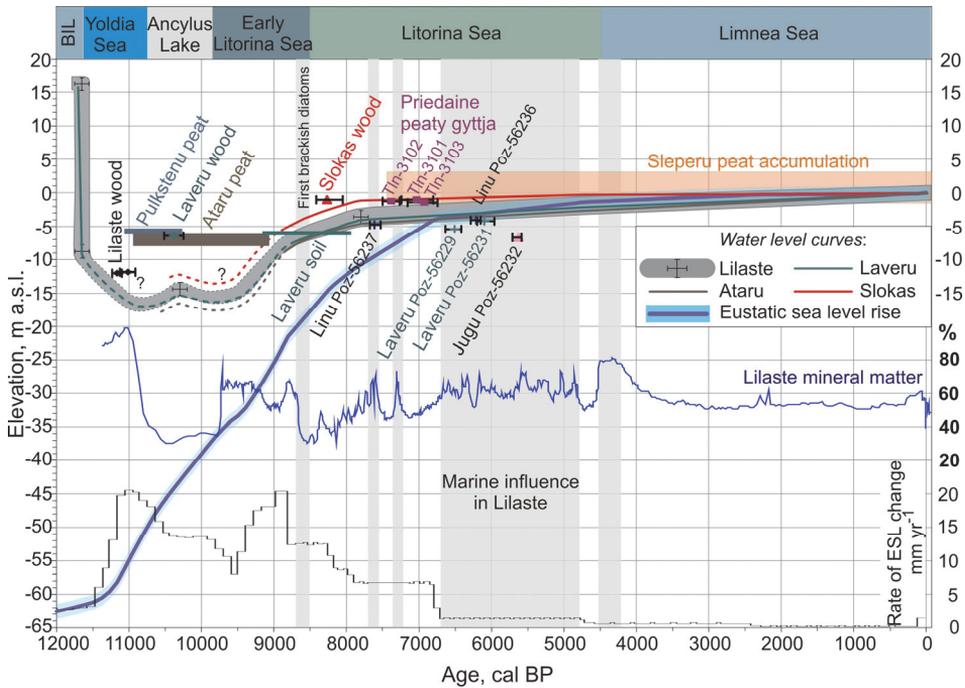


Figure 11. Water-level curves for the southern coast of Gulf of Riga, based on different proxies. Water level for Lake Lilaste is shown with ± 1 m error bar, for lakes Ataru, Laveru and Slokas mean modelled water level is shown. Eustatic sea level data are based on Lambeck et al. (2014).

As proposed earlier by Veinbergs (1996), AL reached 20 m b.s.l.; Kovalenko and Juskevics (1987) reported 5–6 m b.s.l. These assumptions were based mainly on geomorphological and lithostratigraphical evidences with no reliable radiocarbon dates of the shorelines. A low water-level occurrence in the Riga area during the YS phase is indicated by wood remains in Lilaste at 11.9 m b.s.l and 11.72 m b.s.l. dating to 11 200 cal BP and 10 900 cal BP, respectively. Moreover, peat layers were found in lakes Ataru and Pulkstenu at 6.1–7.8 m b.s.l. (9000–10 900 cal BP) and 5.57–5.7 m b.s.l. (10 250–10 970 cal BP), respectively, indicating that a low water level also existed during AL and the Early LitS. Another indication of low water level is the sample of wood at 6.52 m b.s.l. (10380 ± 80 cal BP) and a layer of soil of unknown age at 6.27–6.35 m b.s.l., which were found at the bottom of the sediment sequence of Lake Laveru. The exact time of the formation of the soil cannot be determined because no radiocarbon dates exist from that layer. It should be noted that soil formation can take quite a long period. Research on the pedogenesis in southwestern Estonia in the early Holocene (Reintam et al. 2001) revealed that the soil formation could have lasted for about 1500 years. Preliminary data suggest that the soil formation could have occurred from 9200 ± 800 cal BP to 7900 ± 800 cal BP, as shown in Fig. 11.

The maximum and minimum elevations of RSL during the YS and the AL stages are disputable because the YS and the AL shorelines are under the present sea level in the southern part of Gulf of Riga. Although Veinbergs (1979) reported

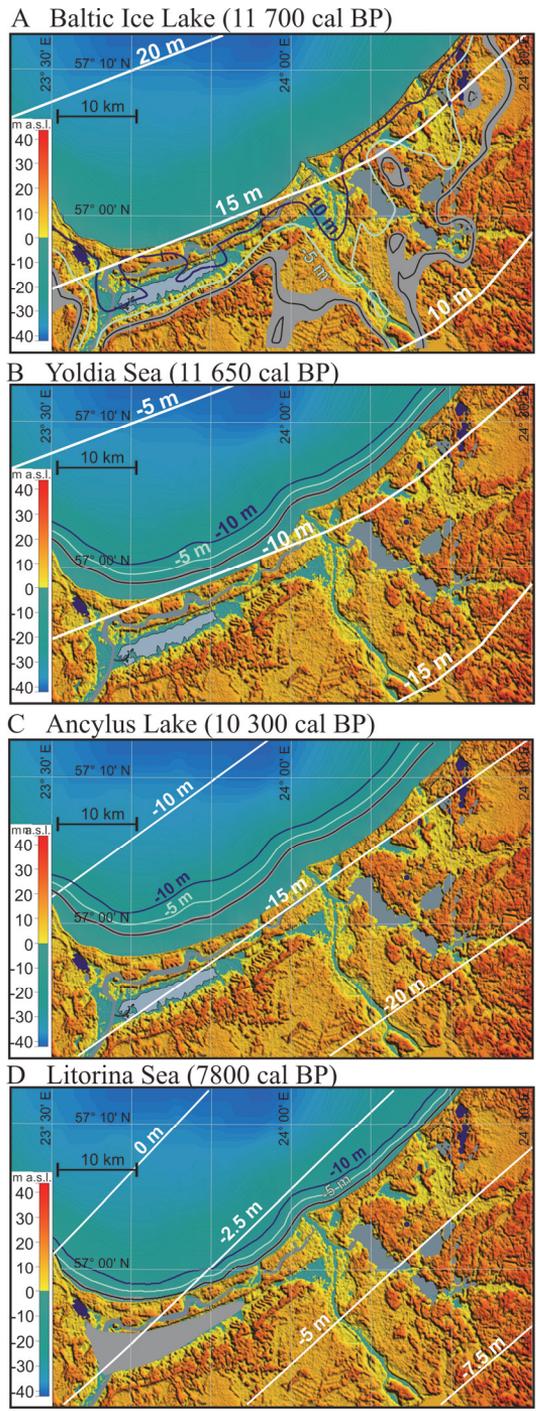


Figure 12. Palaeogeographic maps of the southern coast of the Gulf of Riga. Modelled water level surface isobases are indicated by white lines together with elevations in metres a.s.l. The modelled shoreline is shown with a modelling error $\pm 1\text{m}$ (black line and grey area or grey area with black line); light blue line corresponds to the 5 m and blue to the 10 m water depth.

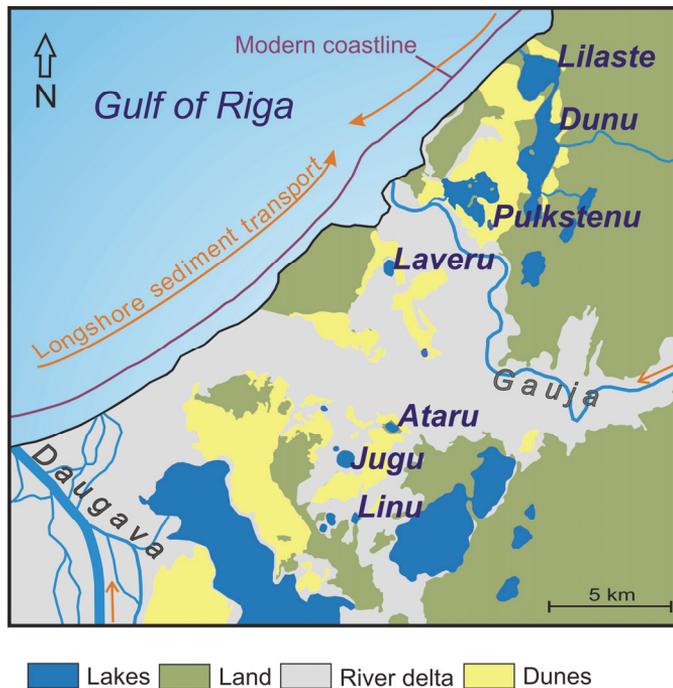


Figure 13. Schematic illustration of area flooded by rivers Daugava and Gauja during the Litorina Sea (6000–5000 cal BP).

that the level of AL was 10–20 m higher than the YS level, an insufficient amount of reliable data prevents determination of the precise amplitude between the YS and AL. The water-level curves obtained in this study, complemented by modelling results of the Baltic Sea shoreline database (Saarse et al. 2003; Vassiljev & Saarse 2013), suggest that the BIL level at 11 700 cal BP was 14–16 m a.s.l., the maximum YS level at 11 650 cal BP was 9–11 m b.s.l., and the maximum of AL transgression at approximately 10 300 cal BP reached 12–17 m b.s.l. in the southern part of the Gulf of Riga (Fig. 12). The low level of the AL supports the sand and gravel deposits ‘Vecdaugava’ (Fig. 2D) occurring 20 m b.s.l. offshore in the southern part of Gulf of Riga in the area between the large Daugava and Gauja rivers. Stelle et al. (1992) proposed that the deposits are a palaeodelta of the Daugava or Gauja formed during the AL.

The reconstructed water-level curves of the southern coast of the Gulf of Riga present RSL increase since the early LitS. A rapid increase in RSL occurred during the early LitS, which could be explained by an increase in the rate of ESL from 8 mm yr^{-1} to 21 mm yr^{-1} at 9500–8800 cal BP (Lambeck et al. 2014). Evidence of rapid RSL rise during the early LitS has been demonstrated in several papers investigating the history of the Baltic Sea (Janke and Lampe 2000; Lampe 2002; Yu 2003) or the water-level increase of the ocean (Lambeck et al. 2014). The sea-level rise caused the groundwater level increase, subsequently resulting in water-level rise in coastal lakes. The first precursors of the LitS transgression at ca 8700 cal BP are brackish water diatoms in Lilaste. The diatom signal of brackish water inflow is weak; therefore, an inconstant connection between

Lilaste and LitS is suggested. Hence, it can be concluded that the level of LitS was lower than the threshold of Lilaste. Prior to the clear evidence of the constant presence of brackish water in Lake Lilaste since 6700 cal BP, it is possible to refer to two distinct sea water surges at ca 7600 and 7300 cal BP that are likely related to strong northwesterly gales that eroded coastal formations. This can explain the peak of brackish-water and halophilous diatoms together with the MM increase at ca 7600 cal BP. This finding is also supported by Yu et al. (2007), who observed a rapid local RSL increase of approximately 4.5 m at about 7600 cal BP in the southeastern Swedish Baltic Sea as response to a sudden increase in ESL likely caused by the final decay of the Labrador sector of the Laurentide Ice Sheet. In addition, a short period of very rapid RSL rise with an amplitude of approximately 4.5 m at 7600–7200 cal BP was detected at the Island of Samsø in Denmark (Sander et al. 2015). Although the ages match, the observed amplitude of the RSL rise in the Riga area was less than that in southeastern Sweden and central Denmark, which could be ≥ 1 m. This result can be concluded from the modelled data showing LitS level of 3.5–4.5 m b.s.l. at ca 7800 cal BP (Fig. 12) at the southern coast of the Gulf of Riga and the estimated height of the threshold of Lilaste at that time. According to the obtained data, the threshold of Lilaste was about 2 m b.s.l., which was reached by the LitS around 6700 cal BP. Therefore, the difference in amplitude of the RSL rise is explained by the disparity in the land uplift rate in these areas.

In the 1950s and 1960s, individual studies were conducted at the southern coast of the Gulf of Riga. Galeniece cited in Grinbergs (1957) analysed a 17-m-long sediment sequence of Lake Kanieris (Fig. 2D) that consisted mainly of carbonaceous gyttja, in which brackish-water diatoms were identified from a depth of 14 m onwards. Another deep sediment core was found in an ancient valley at the western shore of Lake Babites (Fig. 2D). This 16-m-long sediment sequence of dark gyttja with organics was analysed by Aboltina-Presnikova (1960). Peat and sand interlayers at the bottom at 14.7–11 m b.s.l. indicate a shallow water environment, and the presence of brackish-water diatoms at 13–7 m b.s.l. provides evidence of marine water inflows into the Babites basin. Berzin (1967) examined a 35-m-long sediment sequence from the mouth of the Daugava (Fig. 2D). She discovered diatom assemblages in sand with mud at 18.2–14.6 m b.s.l., which indicate a slightly brackish environment. In the overlying mud at 14.6–9.4 m b.s.l., she found brackish-water diatoms, which indicate marine water influence. The authors (Grinbergs 1957; Aboltina-Presnikova 1960; Berzin 1967) propose that sediments with brackish-water diatoms were deposited during the LitS. Unfortunately, this assumption of the age of the studied sediments is based only on pollen zones. A similarity among the aforementioned studies is the depth of appearance of brackish-water diatoms at around 14 m b.s.l. It is worth mentioning that a similar dataset from the southern coast of the Baltic Sea in Germany, the oldest limnic–telmatic sediments overlain by marine deposits were found at approximately 15 m b.s.l. and were dated to around 8500 cal BP (Lampe 2002).

Data from RSL studies at the southern coast of the Baltic Sea indicate that the initial increase in the water table occurred rapidly and reached 2 m b.s.l. around 6800 cal (Lampe 2002). Data from the southern Baltic Sea agree with modelled data from the southern coast of the Gulf of Riga. The reconstructed RSL, correlating with the ESL from 6800 cal BP onwards (Fig. 11), occurred simultaneously with a rapid slowdown of ESL increase from 7 mm yr⁻¹ to 1 mm yr⁻¹ (Lambeck et al. 2014). In the mid-Holocene, ESL rise caused the erosion base of rivers to become higher, transforming the coastal area into a river delta or plain with some small lakes and wetlands filled with fluvial–limnic sediments. Thus, lakes Laveru, Linu, and Jugu (Fig. 13) became part of the river delta. The identified diatom assemblages in these lakes did not confirm LitS intrusion into the river delta. Additional evidence of gradual RSL rise in the Gulf of Riga is provided by peaty gyttja with plant macroremains in the Priedaine lagoon (Fig. 2D) at 1.4–0.9 m b.s.l. at 7400–6900 cal BP, when LitS flooded the depressions and the accumulation of organic sediments began in the lagoon (Eberhards 2008; Kalnina et al. 2009). Gyttja with sand and silt interlayers, as well as the presence of brackish-water diatom species, indicate frequent sea-water inflow into the lagoonal basin. This feature might be related to the temporary water-level increase owing to strong northwesterly winds or storm events.

As previously mentioned, the LitS reached the threshold of Lilaste at ca 6700 cal BP, providing long-term intermittent influxes of brackish water until 4200 cal BP. The distinct brackish-water intrusion from 6700 cal BP to 4200 cal BP coincides with the sea-level stillstand in areas in Norway with low land uplift rates at 6500–4900 cal BP (Balascio et al. 2011). During this time, the transgressive phase of the LitS ended, resulting in the termination of the ESL rise around 6000–5000 cal BP (Eronen 1990). Similar observations of marine water influx during this time were made in the coastal areas of Poland, which were similar to those in Vistula lagoon (Witak and Jankowska 2005) and Szczecin Lagoon (Borówka et al. 2005). The distinct abundance of marine/brackish-water diatom assemblages at 6700–4200 cal BP in Lilaste can be explained by the overall higher salinity in the Baltic Sea during the LitS stage owing to milder and dryer periods with less freshwater run-off (Gustafsson and Westman 2002; Emeis et al. 2003; Zillén et al. 2008).

The brackish-water influence on Lilaste was interrupted for 300 years from 4800 cal BP to 4500 cal BP. The recurrent marine water inflows from 4500 cal BP to 4200 cal BP correlate with the second LitS transgression described by Grinbergs (1957) and Veinbergs (1996). However, the rate of ESL rise decreased from 1 mm yr⁻¹ to 0.5 mm yr⁻¹ around that time (Fig. 11; Lambeck et al. 2014), which does not support the transgression. The water-level curves suggest slow and monotonic RSL rise but does not rule out sea-level oscillations owing to strong westerly/northwesterly wind gusts. Increased storm frequency resulting in erosion of the coastal formations could be one reason for the reappearance of brackish-water diatoms. Additional evidence in conflict with the theory of twofold LitS transgression and limited LitS level less than modern sea level in the southern Gulf of Riga is the Slepere peatbog (Fig. 2D), where wood–reed peat

at 1.5 m b.s.l. began to accumulate at about 7400 cal BP (Apsite et al. 2011; Zaube 2012; Berzins et al. 2012). The dependence of the reed peat surface on the sea level is highly significant and is restricted to an interval of about 0.2 m a.s.l. to 0.2 m b.s.l. (Slobodda 1992, cited after Lampe 2005).

The modelled RSL changes in the southern coast of the Gulf of Riga roughly correlate with the RSL curves from the southern part of the Baltic Sea (Lampe 2005; Uścińowicz 2006). Although both of these areas are close to the zero isobase of uplift, the potential for comparisons between them are limited because the coastal areas of the Baltic Sea are different, with each experiencing unique development regimes. Several factors such as bedrock-forming material, longshore sediment transport, additional riverine material, and coast location from the prevailing winds affect the development of the coast. Changes in RSL during the early Holocene until about 8500 cal BP in the area near the zero uplift, such as the southern coast of the Gulf of Riga and the coastal areas of northern Poland and northeastern Germany, were related primarily to deglaciation dynamics (Uścińowicz 2006). The rate of ESL increase reduced at 8200–6700 cal BP, which is consistent with the final phase of North American deglaciation (Lambeck et al. 2014). The sea-level curve of the southern coast of the Baltic Sea revealed that the level of 1–1.5 m b.s.l. had been reached by 5800 cal BP and that the sea level rose to about 0.6 m b.s.l. until 4000 cal BP (Lampe 2005). Based on the modelled data, at about 4800 cal BP the sea level was 1.5 m b.s.l. in the southeastern, and 0.5 m b.s.l. in the southwestern part of the Gulf of Riga.

Lampe (2002) reported that at least in the past 5000 years, the RSL increased on a rather smooth course and was interrupted by few minor fluctuations. Since the end of the main LitS transgression, neither periodical sea-level oscillations nor fluctuations in the range of 1 m or more could be detected in the deposits; the opposite was at times reported previously (Lampe 2005). The total ESL increase for the past 6700 years was approximately 4 m, of which about 3 m occurred at 6700–4200 cal BP with a further rise of ≥ 1 m up to the onset of recent sea-level rise approximately 100–150 years ago (Lambeck et al. 2014). In this interval of 4200 cal BP to around 150 years ago, no evidence was detected for oscillations in ESL of amplitudes exceeding 0.15–0.20 m on time scales of about 200 years (Lambeck et al. 2014).

In accordance with observations in northwestern Estonia (Papers I–III), southern Finland (Seppä et al. 2000; Eronen et al. 2001; Tikkanen & Oksanen 2002; Miettinen 2004), the Island of Hiiumaa (Paper IV), the southern coast of the Gulf of Riga (Paper V), and the southern coast of the Baltic Sea (Lampe 2005; Uścińowicz 2006), the diatom compositions do not support the concept of the twofold LitS transgression suggested earlier. These results do support one main transgression event, however. The high-amplitude RSL fluctuations require forcing mechanisms or climate changes that are currently unknown (Baeteman et al. 2011). Minor temporary fluctuations are normal and can be correlated with climate oscillations likely caused by changes in the North Atlantic circulation (Lampe 2005) or can be explained by increased storminess (Baeteman et al. 2011). Temporary fluctuations in the RSL during strong northwesterly winds or

storm surges can be currently observed (Eberhards 2003). During the late Holocene, active longshore sediment transport and additional MM from rivers contributed to the formations of bars, foredunes, spits, and wider beaches. Presently, the heights of foredunes in the study area can reach up to 6 m a.s.l. However, inner dunes up to 20 m a.s.l. (Paper V) currently protect the Lilaste from storm surges and marine water intrusion into the lake.

6. CONCLUSIONS

The thesis provides insight into the history of post-glacial changes of RSL and illustrates the pattern of land uplift in the eastern part of the Baltic Sea by combining data from three different study sites: northern and northeastern Estonia, the Island of Hiiumaa, and the southern coast of the Gulf of Riga. This study marks the first attempt to reconstruct water-level curves for a wide variety of settings based on high-resolution bio-, litho-, and chronostratigraphical evidence from sediment records of isolation basins in Latvia.

The chosen multi-proxy approach that includes diatom analysis proved to be an effective tool in identifying the position of the isolation contact in the sediment sequence and in defining the related changes in basin salinity and isolation dynamics. On the basis of the obtained results and GIS analysis, palaeogeographical maps for various time windows were compiled, which together with palaeogeographical descriptions can serve as an effective tool for archaeologists in examining new prehistorical settlements.

In the area with higher land uplift, the LitS transgression occurred earlier such as on the Island of Hiiumaa at 8800–8200 cal BP and in northern Estonia at 7800 cal BP. In the areas close to zero isoline of apparent land uplift, however, the LitS level rose gradually, representing the global pattern of ESL changes. The LitS reached a level close to the present sea level at 5000–4200 cal BP at the southern coast of the Gulf of Riga.

The RSL in northern and northwestern Estonia and on the Island of Hiiumaa was regressive after the LitS transgression and throughout the LimS stage because the land uplift rate exceeded the ESL increase. The opposite occurred at the southern coast of the Gulf of Riga, where the water level rose gradually along with changes in ESL. No evidence of a twofold or multiple transgression of the LitS was indicated in the eastern part of the Baltic Sea.

The isolation of basins in northern Estonia and the Island of Hiiumaa was dependent mainly on the land uplift rate as it exceeded the ESL rise. However, at the southern coast of Gulf of Riga, where the land uplift was nearly zero, the process of isolation was influenced by gradual ESL rise, river delta infilling by sediments, long-shore sediment transport, and the formation of new beach areas.

The reconstructed water-level curves in northern Estonia and on the Island of Hiiumaa show the same trend of sea-level changes as the recently reconstructed curves around the Gulf of Finland and the Gulf of Bothnia. The reconstructed relative water-level curves in the southern Gulf of Riga correlate with the ESL from 6800 cal BP onwards and exhibits the same RSL pattern as the curves from the southern Baltic Sea.

The water level in the Baltic Sea basin until 8500 cal BP was influenced primarily by deglaciation dynamics, whereas in the last 8500 years, the main factor was complicated interplay between the ESL increase and the pattern of land uplift.

The strong point of this thesis is the multi-proxy approach by applying high-resolution studies of wide settings of isolation basins to obtain new evidence of RSL changes and patterns of land uplift. This study can serve as the basis for further investigations in other areas along the coastline of the eastern Baltic Sea. Moreover, it may help to provide answers to unclear questions in the future and resolve problems concerning postglacial land uplift and changes in RSL in this area.

Northern and northwestern Estonia, the Island of Hiiumaa, and the southern part of the Gulf of Riga demonstrate similarities of the main pattern of the Baltic Sea history including BIL drainage, YS regression, AL transgression and regression, and rapid rise of the LitS level. However, it should be considered that each of these areas experienced a unique regime of glacio–isostatic rebound and RSL change.

7. ABSTRACT

The main objectives of this thesis are to reconstruct changes of relative sea level (RSL) and to determine the pattern of land uplift in the eastern Baltic over the Holocene. The reconstruction is based on isolation basin studies determined after the isolation contact in the isolation basins, the height of the threshold, and geomorphological markers. Other issues discussed in this thesis include the number of Litorina Sea (LitS) transgressions in the eastern Baltic Sea, credibility of the data, and the number of study sites in areas with complicated sediment stratigraphy.

Lake sediment sequences from five lakes in northern and northwestern Estonia, three from the Island of Hiiumaa and eight from the southern coast of the Gulf of Riga were analysed by using multi-proxy studies such as diatom analysis, loss-on-ignition (LOI), magnetic susceptibility (MS), grain size analysis, Accelerator Mass Spectrometry (AMS) and conventional ^{14}C datings. Additionally, spheroidal fly ash particles were used to date upper loose sediments of Lake Lilaste. Finally, palaeogeographical reconstructions were based on GIS analysis in which interpolated water-level surfaces were removed from the digital terrain model (DTM).

The obtained data were applied for reconstructions of RSL curves that enabled the detection of water-level fluctuations and patterns of land uplift in northern and northwestern Estonia, the Island of Hiiumaa, and the southern coast of the Gulf of Riga. Water-level curves from northern Estonia and the Island of Hiiumaa show a smooth decrease in RSL and land uplift, which have been close to linear since the mid-Holocene. Data from the surrounding areas of Riga confirm that the LitS transgressed consistently at the southern coast of Gulf of Riga, representing a pattern of eustatic sea level (ESL) changes.

The reconstructed water-level curves display a diachronous LitS transgression peak in the Island of Hiiumaa that occurred earlier in the Kõivasoo area at about 8800 cal BP and later in the Prassi area at approximately 8200 cal BP (Paper IV). The LitS maximum in northern Estonia at about 7800 cal BP (Paper II) was later than that at the Island of Hiiumaa. The time-transgressive LitS transgression was observed in the Baltic Sea region as a result of the different rates of land uplift. The LitS transgression culminated almost 1000 years later and was a long-lasting event (about 2500 years) in the southern coast of the Gulf of Riga (Paper V) compared with sites at high isolines in the northern part of the Baltic Sea.

The calculated apparent land uplift during the LitS transgression in the Island of Hiiumaa was 3.4 mm yr^{-1} at Kõivasoo and 2.8 mm yr^{-1} at Prassi and has decreased to 2.5 and 2.1 mm yr^{-1} at present (Paper IV). Similar declines of land uplift are observed in northern and northwestern Estonia. The uplift rate during LitS transgression was approximately 2.8 mm yr^{-1} in the area of Tānavjārv, which has decreased to 2.2 mm yr^{-1} at present, and at Lohja from 2.4 mm yr^{-1} to about 2.0 mm yr^{-1} (Paper II). However, preliminary calculations of the apparent uplift at Lilaste in the southern coast of Gulf of Riga reveals that this area experienced minor subsidence of about -0.8 mm yr^{-1} between 6500 cal BP and 4600 cal BP,

and during the last 4600 years, the apparent land uplift rate has been near -0.1 mm yr^{-1} .

Lilaste and other lakes in the surrounding areas of Riga indicate a complicated sediment deposition and changes in RSL over the Holocene in the area with the apparent land uplift close to 0 mm yr^{-1} . The sediment sequences of the studied lakes contain evidence of low water-levels, such as peat/soil layers and buried wood, during the Yoldia Sea (YS), Ancylus Lake (AL), and Early LitS stages, as well as slow water-level increases, river bifurcation, and intermittent LitS brackish-water inflows that promoted nutrient enrichment in the isolation basins (Paper V). In only two of the eight studied basins Lilaste and Slokas, the diatom analysis confirmed that the LitS reached the threshold and influenced their environment. According to the pattern of the reconstructed water-level curves, the RSL of the Baltic Sea in the early Holocene was related primarily to deglaciation dynamics, although since about 8500 cal BP, the key role has been the interplay between glacio–isostasy and ESL rise.

Similarities in diatom assemblages were observed in the studied isolation basins, such as the mass occurrence of small-sized *Fragilaria* spp. and increased nutrient content. The mass occurrence of small-sized fragilarioid taxa just before, during, or after the isolation has been recognised by several authors in different studies. *Fragilaria* spp. indicates a period of environmental instability due to increased turbidity and higher nutrient content (Paper I–V). The peak of *Stephanodiscus parvus* and *Cyclotella meneghiniana*, indicating an increased nutrient content and a high conductivity state, are identified in the basins in which isolation from the sea was a long-lasting event (Paper I–III, V). The enhanced nutrient content might be explained by occasional mixing of brackish water and fresh water, which promotes biological productivity.

This thesis demonstrates the value of the multi-proxy approach in a wide number of settings, particularly the use of diatom analysis, to determine the palaeoenvironmental changes and to reconstruct the RSL. The obtained data and conclusions contribute to earlier knowledge of water-level change and shoreline displacement in the eastern Baltics. The approach used to study the southern coast of the Gulf of Riga, including well-dated Lake Lilaste, can be used as the basis for further investigation in other areas along the Latvian coastline.

8. KOKKUVÕTE

Ränivetikate stratigraafia ja Läänemere idaosa veetaseme muutused Holotseenis Väitekiri käsitleb Läänemere idaosa pärastjääaegse rannajoone siirdeid, glatsioisostaatilise maakerke iseärasusi ja rannikumere soolsuse muutuseid. Uurimispiirkond hõlmab Eesti põhja- ja loodeosa, Hiiumaa ning Riia lahe lõunarannikut. Biostratigraafiliste meetoditega (ränivetikad, õietolm) ning litostratigraafiliste parameetritega (setete kuumutuskadu, magnetiline vastuvõtlikkus, lõimis) määrati eri kõrgusel asetsevate settebasseinide läbilõigete isolatsiooni tasemed. AMS radiosüsiniku dateeringute põhjal määrati uuritavate settebasseinide isoleerumise aeg tagades nii rannajoone rekonstruktsioonidele täpse ajaskaala. Tuginedes GIS ja DMT mudelitele rekonstrueeriti Läänemere erinevate staadiumite rannajooned ning kolmemõõtmelised paleogeograafilised kaardid. Läti ala kohta on tegemist esmakordse sellise detailsuga uuringuga.

Uuringuala arvutuslik suhteline glatsioisostaatilise maakerge on suurim Hiiumaal, Litoriinamere transgressiooni ajal tõusis maapind Kõpus 3,4 mm aastas ja Prassis 2,8 mm aastas, tänapäeval on maakerge tunduvalt väiksem, vastavalt Kõpus 2,5 mm aastas ja Prassis 2,1 mm aastas. Sarnane suhtelise maakerke kiiruse vähenemine on omane ka Eesti loode- ja põhjaosale. Kui Litoriinamere transgressiooni ajal tõusis maapind Tānavjärve ümbruses 2,8 mm aastas ja Lohjal 2,4 mm aastas, siis tänapäevaks on maakerge kahanenud Tānavjärve ümbruses 2,2 mm aastas ja Lohjal 2,0 mm aastas. Riia lahe lõunarannikul on maapind vajunud. Ajavahemikul 6500 ja 4600 aastat tagasi on arvutuslik suhteline maa vajumiskiirus 0,8 mm aastas, kuni tänapäevani on maapinna alanemise kiirus püsinud 0,1 mm aastas.

Rannasiirdekõverad kinnitavad soolaseveelise Litoriinamere transgressiooni maksimumi eriaegsust. Hiiumaal Kõpu poolsaarel Kõivasoo lähistel oli Litoriinamere veetase maksimaalne 8800 aastat tagasi, seevastu Lõuna-Hiiumaal Prassis 8200 aastat tagasi. Litoriinamere transgressiooni maksimum Põhja-Eestis oli veelgi noorem, 7800 aastat tagasi. Litoriinamere transgressiooni maksimumi eriaegsust Läänemere idaosas põhjustas erinev maakerke kiirus. Litoriinamere transgressiooni maksimum Riia lahe lõunarannikul on veelgi noorem ja pikemaajalisem protsess, mis kestis ajavahemikul 6700–4200 aastat tagasi. Erinevalt Läänemere lääneosast, kus Litoriinamere arengus esines mitu transgressiooni, esines Läänemere idaosas vaid üks meretaseme tõus. Kuna Eesti uurimisalal glatsioisostaatiline maakerge ületas maailmamere eustaatilist tõusu, siis järgnes Eestis Litoriinamere aegsele transgressioonile meretaseme ühtlane langus. Seevastu Läti alal rannasiirdekõver järgib globaalset ookeani veetaseme eustaatilist tõusukõverat.

9. ACKNOWLEDGEMENTS

I wish to thank the following people, without whom this work would not have been possible. First and foremost, I wish to express my special appreciation and thanks to both of my supervisors, Siim Veski and Atko Heinsalu. I would like to thank Siim Veski for his patience, motivation, positive outlook, and immense knowledge. I have been given a unique opportunity to join his research group. I want to thank Atko Heinsalu, who has initiated this work and contributed with many discussions concerning diatom identification, grouping, and results interpretation.

Similarly, profound gratitude goes to co-authors Leili Saarse and Jüri Vassiljev. I want to thank Leili for her tremendous academic support and for the continuous help in my PhD study and Jüri for his contribution of knowledge in modelling water-level curves, creating figures of palaeoreconstructions, and valuable comments about interpreting that data. Moreover, I would like to thank my other colleagues including Tiiu Alliksaar for explaining the grain size and spheroidal fly ash particle methods that I applied in my research. I thank my fellows Merlin Liiv and Normunds Stivrins for being the best roommates ever. I have very fond memories for all the fun we have had in the last four years and, of course, our excursions in which we became acquainted with the most beautiful places in Estonia. Now I can truly say that we do not need words to understand each other. Particular thanks go to Normunds for nurturing my enthusiasm for geology; in other words, thank you for being my scientific soul mate.

My sincere thanks also go to Jan Risberg and Kaarina Weckström, who provided me with an opportunity to gain experience in their institutions and who granted access to the laboratories and research facilities. Jan was the first to introduce me to the diatom world in my master's studies. I am also hugely appreciative to one of the most prominent geologists in Latvia, Guntis Eberhards, for his time allocated to consultation about shoreline research and coastal processes in Latvia.

My thanks are due to the Mafia of Geographers (Ģeogrāfu mafija) or, in other words, my colleagues and friends from Latvia. During the first field course in geology Laimdota Kalnina lured me into a paleoecology trap as far back as 2006. Thanks to Aija Cerina for sharing her vast experience in palaeoecology. I would like to thank Eliza Kuske, Ilze Ozola, Agnese Pujate, Vita Ratniece, and Sandra Zeimule being my Latvian support team throughout this study period. In addition, thanks go to Agnis Recs for helping in field work by measuring the exact altitude, to Toms Bricis for weather forecasting and consultations prior to the field work, and to Janis Bikse for providing me with useful maps and help with computer issues.

Moreover, I thank my colleagues from the University of Tartu. Thanks go to Alar Rosentau for offering useful courses and field praxis concerning shoreline displacement and palaeoreconstruction in addition to pleasant co-operation. Thanks also go to Hanna Raig for stimulating discussions about diatom

identification and their separation into ecological groups. Thank for your kind interest in my work.

I would also like to thank all of my friends who supported me in writing and for encouraging me to strive towards my goal. I am particularly indebted to Ilze Stundina for her constant faith in my work and for her support by generously hosting me in Riga during my trips from Tallinn to home. I would like to express appreciation to Edite Kulmane for supporting me by intimating the right words while I was away from home, friends, and family. I would like to thank Vladislavs Grekis for encouraging me to undertake doctoral studies in Tallinn and to always move forward to achieve my goals.

Special thanks go to my loved ones. Words cannot express my gratitude to my family for supporting me spiritually throughout the study years. Finally, I would like to greatly thank Armands for providing almost unbelievable support; he always gave me support when no one was available to answer my queries.

I gratefully acknowledge the financial support from the Doctoral Studies and Internationalisation Programme DoRa for giving me the opportunity to study at TTU GI, SPLASHCOS for financing my study visit at GEUS in Denmark, Maria Kahlert for providing participation in NORBAF, and the Doctoral School of Earth Sciences and Ecology for financing my participation in conferences, workshops, and seminars.

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Saksa keel	kesktase
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APPENDIX: PUBLICATIONS

I Grudzinska, I., Saarse, L., Vassiljev, J., Heinsalu, A., Veski, S. 2012. A palaeocoastline reconstruction for the Käsnu and Pärисpea peninsulas (northern Estonia) over the last 4000 years. *Estonian Journal of Earth Sciences* 61: 307–316. doi: 10.3176/earth.2012.4.09

A palaeocoastline reconstruction for the Käsü and Pärissa peninsula (northern Estonia) over the last 4000 years

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Received 29 June 2012, accepted 28 September 2012

Abstract. The marine-freshwater environmental transition, i.e. basin isolation from the Limnea Sea, has been identified in two short sediment cores with respect to their diatom composition, loss-on-ignition and magnetic susceptibility content. The isolation level of the basins was dated by accelerator mass spectrometry radiocarbon measurements. The basins are situated on the southern coast of the Gulf of Finland between altitudes 5.5 and 3.9 m above sea level. The Lohja basin became isolated from the sea around 2300 cal yr BP and Käsü basin around 1800 cal yr BP as a result of glacio-isostatic uplift. The total land upheaval rate that has been 2.6 mm yr⁻¹ since 2500 cal yr BP has currently decreased to 2.0 mm yr⁻¹. We present a GIS-based 3D palaeogeographic reconstruction of the palaeocoastline changes in northern Estonia for two peninsulas, Pärissa and Käsü, as well as compose a shoreline displacement curve for the study area, which is a compilation of previous and ongoing investigations.

Key words: shoreline displacement, palaeogeography, Limnea Sea, diatoms, isostatic land uplift, Estonia.

INTRODUCTION

The North Estonian coastline is highly jointed. The post-glacial palaeocoastline formation in the Baltic Sea basin has been ruled by the deglaciation of the Fennoscandian Ice Sheet and resulting isostatic rebound together with eustatic sea level changes. Several peninsulas stretch far out to the sea and bays invade the mainland. These peninsulas are young, formed during the Holocene, when islands started to emerge in the offshore areas of the Gulf of Finland, to become joined with the mainland at a later point and form peninsulas and a winding coastline. Studies have pursued to define the pattern of post-glacial shoreline displacement of northern Estonia since the early 20th century (e.g. Ramsay 1929; Kents 1939; Kessel & Raukas 1979; Hyvärinen et al. 1988). Contemporary research that has endeavoured to reconstruct the palaeogeographic development of the coastline changes in northern Estonia has centred on the early stages of the Baltic Sea basin, i.e. the development of the Baltic Ice Lake (Rosentau et al. 2007; Vassiljev et al. 2011), Yoldia Sea (Heinsalu & Veski 2007), Ancylus Lake (Veski 1998; Saarse et al. 1999) and Litorina Sea (Saarse et al. 2009, 2010). In general, little is known about the Late Holocene shoreline displacement patterns on the Estonian coast of the Gulf of Finland. Due to lack of dated isolation sequences at lower altitudes, relative sea-level changes for the Limnea Sea are poorly constrained and the lower sections of the shoreline displacement curves are mainly based on extrapolation.

The main objective of the study is to reconstruct palaeogeography for two North Estonian peninsulas, Pärissa and Käsü. Beach formations of these peninsulas have been examined and levelled earlier (Linkrus 1969, 1971); however, their age has remained uncertain due to lack of radiocarbon dates. In the current study diatoms, loss-on-ignition (LOI), magnetic susceptibility (MS) and radiocarbon dates of the sediment cores of two North Estonian lakes have been examined for the purpose of detecting the age of the isolation of these lakes from the Limnea Sea, acting as a complement to older investigations (Kessel & Linkrus 1979; Heinsalu 2000; Saarse et al. 2006, 2009, 2010; Saarse & Vassiljev 2010) in compiling a new shore displacement curve for the area. Such a methodology has been widely used in relative sea level reconstructions (e.g. Snyder et al. 1997; Miettinen et al. 2007; Long et al. 2009; Watcham et al. 2011; Lunkka et al. 2012; Saarnisto 2012) as well as in studies of isostatic uplift (Risberg et al. 1996).

STUDY AREA

The studied lakes Lohja and Käsü are located in the North Estonian coastal lowland (Fig. 1), on the terraces of the Limnea Sea. The lakes occupy depressions separated from the sea by beach ridges and dunes and their catchment is mostly covered by sand. According to the water level simulation, the highest shoreline of the Litorina Sea reached 18.8 m a.s.l. at Lohja and 17.7 m



Fig. 1. Location of the study area, marked by a square.

a.s.l. at Käsü and that of the Limnea Sea 10.7 m and 10.2 m, respectively. The modern land uplift relative to sea level is 2 mm yr^{-1} at Lohja and between 2 and 1.5 mm yr^{-1} at Käsü (Torim 2004). Prior to isolation, the Lohja basin was located in the rump of Hara Bay that was opened on the northwestern side of the lake, whereas the Käsü basin formed a lagoon with two narrow passages in the east.

Lake Lohja (sediment core $59^{\circ}32'57''\text{N}$, $25^{\circ}41'23''\text{E}$, water depth 2.4 m) is located in the southwestern corner of the Pärисpea Peninsula, between the klint escarpment and the sea coast at an elevation of 5.5 m a.s.l. It is a small shallow lake with a surface area of 56 ha, average depth of 2.2 m and maximum depth of 3.7 m (Riikoja 1934; Mäemets 1977). A number of beach ridges and dunes surround the lake in the north and west. Lake Lohja is a dark-coloured soft-water lake with outflow via Lohja Brook to Hara Bay. The water table of the lake is artificially regulated by a dam. The lake has low water transparency due to its brown water (Secchi disc depth $< 1 \text{ m}$). The total phosphorus (TP) and total nitrogen (TN) concentrations are relatively high, $40 \mu\text{g L}^{-1}$ and $700\text{--}900 \mu\text{g L}^{-1}$, respectively.

Lake Käsü (sediment core $59^{\circ}34'56''\text{N}$, $25^{\circ}52'51''\text{E}$, water depth 2.4 m), with an area of 48 ha and maximum water depth 3 m, lies on the Käsü Peninsula 200 m from the sea coast at an altitude of 3.9 m a.s.l. A small outflowing brook joins it with Käsü Bay. The lake level has been regulated. The ditch west to Eru Bay lowered the water level by almost 1 m, but a rebuilt dam on the ditch elevated the water back to its previous level. The lake is boarded by beach ridges in the west and by a flat slightly paludified marine plain in the east, covered by different types of forest. Lake Käsü is a dark-coloured soft-water lake, characterized by high concentrations of TP, $50 \mu\text{g L}^{-1}$,

and TN, $750 \mu\text{g L}^{-1}$. The lake water is rich in organic compounds and poor in total dissolved solids.

MATERIALS AND METHODS

Four overlapping cores were obtained in winter 2011 with a Russian peat sampler from the deepest part of the basins. One metre long core sections were described on site, photographed, sealed in plastic half tubes, transported to the laboratory and stored in a cold-room. Continuous 1 cm thick samples were used for LOI analyses. The organic matter (OM) content was measured by LOI at 525°C for 4 h and expressed in percentages of dry matter. The percentage of carbonates (CaCO_3) was calculated after combustion of LOI residue for 2 h at 900°C . The amount of residue containing terrigenous matter and biogenic silica was described as mineral matter and calculated from the sum of the organic and carbonate compounds. Magnetic susceptibility was measured with a Bartington MS2E high-resolution scanning sensor. The sediment surface was cleaned, covered with a thin plastic film and MS was measured from the sediment surface at 1-cm resolution.

Diatom analyses from Lake Käsü were made by E. Vishnevskaya several years ago (Kessel et al. 1986). For this reason diatom taxonomy was modified, diagram redrawn and sediment sequence correlated with the new one on the basis of lithostratigraphy and LOI results. The difference between the sediment thickness of the previous and the current core accounts for approximately 20 cm, which has been considered in the correlation of the cores. The diatom preparation was carried out, following the techniques outlined in Battarbee (1986). Diatom samples were digested in hydrogen peroxide until the organic matter was oxidized, followed by removing fine and coarse mineral particles by repeated decantation. A few drops of the remaining residue were dried onto cover slips and permanently mounted onto microscope slides, using Naphrax resin. At least 400 diatom valves were identified and counted from each subsample under Zeiss Axio Imager A1 microscope at $\times 1000$ magnification, using oil immersion and differential interference contrast optics. Diatoms were grouped according to their habitat into plankton and periphyton, and according to their salinity tolerance, to brackish/marine, halophilous and freshwater taxa. Diatom and LOI results were plotted, using the TGView software (Grimm 2007).

Macrofossils for radiocarbon dating were extracted by soaking 1 cm thick samples in water and $\text{Na}_4\text{P}_2\text{O}_7$ solution and then sieving the material through a 0.25 mm mesh. Macrofossil remains were identified under a binocular microscope. Carefully selected terrestrial material was dated in the Poznan Radiocarbon Laboratory. Radiocarbon dates (Table 1) were calibrated at one-

Table 1. AMS radiocarbon dates. The ages have been calibrated according to Reimer et al. (2009)

Lake	Basin altitude, m a.s.l.	Depth, cm	Age, ^{14}C yr BP	Age, cal yr BP with midpoint in brackets	Lab. No.	Dated material
Lohja	5.8	365–370	2275 ± 30	2185–2345 (2270 ± 80)	Poz-42171	<i>Pinus</i> bark
Lohja	5.8	395	2490 ± 35	2490–2710 (2600 ± 110)	Poz-42172	<i>Tilia</i> wood
Käsmu	3.9	416	1830 ± 30	1730–1815 (1770 ± 40)	Poz-42177	<i>Pinus</i> bark
Käsmu	3.9	433	1910 ± 30	1825–1885 (1850 ± 30)	Poz-42166	<i>Pinus</i> wood

confidence level, using the Calib Rev 6.0.1 software (Reimer et al. 2009). Alternative chronology was created with an OxCal deposition model (Bronk Ramsey 2008, 2009), where AMS dates were combined with lithological boundaries. Both chronologies differ only slightly and therefore we present here Calib software results. Calibrated ages before present (cal yr BP; 0 = AD 1950) were applied in the present study. Due to the low level of salinity in the Gulf of Finland (Eronen et al. 2001; Miettinen et al. 2007), the marine calibration set was not considered.

Palaeogeographical maps were reconstructed, using GIS techniques. The interpolated surfaces of water level were derived from the Baltic Sea shoreline database (Saarse et al. 2003, 2006) and earlier published sources (Linkrus 1976, 1988), using a point kriging approach. Topographic maps at a scale of 1 : 10 000 and 1 : 25 000 were used to create a digital terrain model (DTM) with grid size 20×20 m. Holocene peat deposits were removed from the DTM. Series of palaeogeographical maps were compiled based on the assumption that during the last 5000 years land uplift diminished linearly (Mörner 1979; Lindén et al. 2006; Yu et al. 2007) and global sea level remained nearly constant (Lambeck & Chappell 2001).

RESULTS

A 195 cm thick sediment core was taken from the central part of Lake Lohja. The basal grey silt (435–375 cm, Lo-1, Fig. 2A) is overlain by greenish-brown silty gyttja (375–365 cm, Lo-2) and dark brown gyttja (365–332 cm, Lo-3). The OM and carbonate content in the silt is low, in silty gyttja OM increases from 10% to 20%. In the lowermost portion of silt (435–390 cm) MS is uniform, about $5\text{--}6 \times 10^{-5}$ SI units; between core depths of 390 and 355 cm MS continuously decreases and after that stabilizes, being in good accord with the mineral matter inclination (Fig. 2A). A distinct lithostratigraphical

change occurs in between 375 and 365 cm, where the OM content apparently rises and stabilizes afterwards.

The lower part of the core (400–365 cm) contains abundant benthic brackish/marine diatom flora, notably *Planothidium delicatulum* and *Navicula peregrina* (Fig. 3A). Simultaneously, small-sized fragilarioid epipsammic diatoms with brackish-water affinity, such as *Opephora mutabilis*, *O. guenter-grassi* and *Fragilaria gedanensis*, occur with high values, accompanied by freshwater *Pseudostaurosira elliptica*. The absolute dominance of periphytic diatoms suggests shallow-water conditions in the sheltered lagoon-like basin that was connected with the Gulf of Finland through the open strait. The diatom composition shows a relatively abrupt transition from brackish/marine diatom flora to predominantly freshwater species at a core depth of 365 cm, indicating a relatively abrupt isolation contact. The planktonic freshwater diatom *Aulacoseira subarctica* predominates together with *A. ambigua* and *A. granulata*. An AMS radiocarbon date of 2270 ± 80 cal yr BP (Poz-42171, Table 1) obtained from a pine bark fragment represents an approximate date of the isolation of the basin.

Lake Käsmu was cored in its western part, where a 360 cm thick sediment sequence was recovered. The basal sand (600–595 cm, Kä-1), overlain by a silt layer (595–433 cm, Kä-2 and Kä-4) and comprising a thin sand layer at 495–475 cm (Kä-3), is poor in OM (Fig. 2B). From 435 cm upwards the content of OM gradually increases to 35% and of carbonates to 5%. The MS curve is changeable. It has decreased considerably since the isolation of the basin as supply of mineral matter has reduced (Risberg et al. 1996).

The diatom assemblage at the bottom of the core contains abundant taxa with marine and brackish-water affinity (Fig. 3B). Pelagic forms like *Chaetoceros holsaticus*, *Pauliella taeniata* and *Thalassiosira baltica*, as well as periphytic taxa such as *Cocconeis scutellum* and *Tabularia fasciculata* are common, indicating rather open bay-like conditions with brackish-water environment. At a core

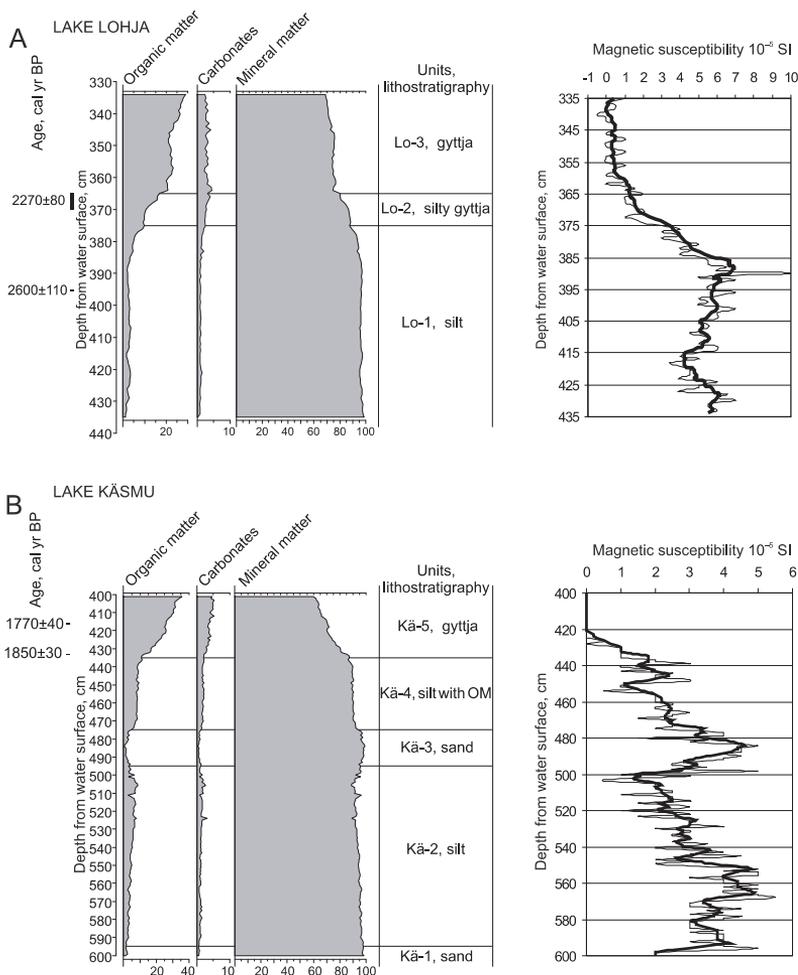


Fig. 2. Loss-on-ignition and magnetic susceptibility of Lake Lohja (A) and Lake Käsmu (B). The thick black line in the magnetic susceptibility graph is 5-sample moving average.

depth of 500 cm, relative abundance of brackish-water diatoms decreases and that of pelagic diatoms and small-sized fragilarioid taxa increases significantly, which suggests the lowering of water depth and the formation of a shallow sheltered coastal lagoon that was connected with Käsmu Bay through narrow channels. Relative abundance of brackish-water diatoms decreases at 430 cm. These are replaced by periphytic halophilous species and by planktonic *Cyclotella meneghiniana*. *Cyclotella meneghiniana* grows in variable environmental conditions – in brackish waters with elevated nutrient concentration (Weckström & Juggins 2006), lakes with high conductivity

(Saros & Fritz 2000) or hypereutrophic lakes (Bradshaw et al. 2002). Diatom-derived isolation is located at a depth of 400 cm, above which the level of brackish-water diatoms has declined and freshwater planktonic diatoms such as *Aulacoseira subarctica*, *A. ambigua* and *A. granulata* dominate, reflecting the development of a small eutrophic lake. An AMS ¹⁴C date of the pine bark fragment from a depth of 416 cm yielded an age of 1770±40 cal yr BP (Poz-42177, Table 1), corresponding to a time shortly before the isolation. Therefore the final isolation of Lake Käsmu took place approximately 1800–1700 cal yr BP.

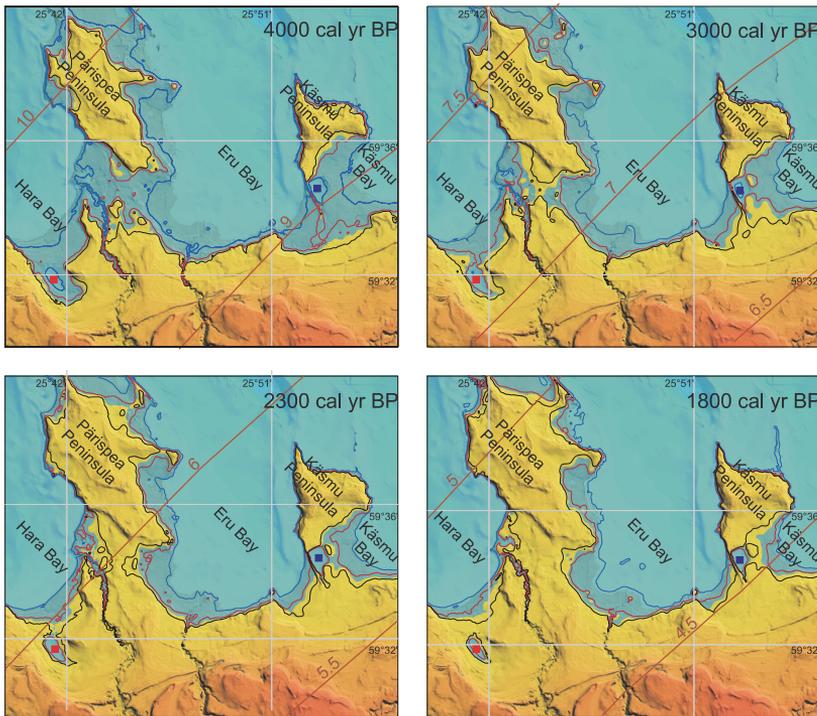


Fig. 4. Palaeogeographic maps for time windows 4000, 3000, 2300 and 1800 cal yr BP. Modelled water level surface isobases are indicated by brown lines together with altitudes in metres a.s.l. The shoreline location is shown together with a possible modelling error in the range of ± 1 m (black line +1 m, red line -1 m); the blue line corresponds to the modelled water depth of 5 m. The reconstructions are overlaid by LIDAR elevation data (Estonian Land Board) to visualize the location of the present-day land (greenish) and sea (blue). Lake Lohja is shown by a red square, Lake Käsnu by a blue square. Reconstruction by J. Vassiljev.

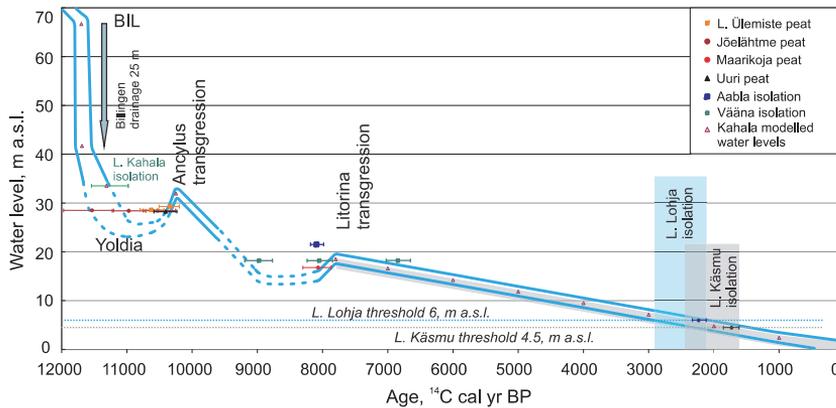


Fig. 5. Shoreline displacement curve for the study area. For the Lohja site blue lines mark the modelled possible minimum and maximum water level changes and for the Käsnu site, the grey area (shown only from the Litorina Sea transgression up to today) marks the modelled possible range of the water level. The reconstruction considers errors in both, modelled water levels (± 1 m) and ages. The blue and grey boxes show the possible isolation age according to modelled water level against the isolation age according to the AMS ^{14}C dates (black circles with error bars).

DISCUSSION

A GIS-based palaeogeographical reconstruction visualizes temporal coastline development in northern Estonia during the last 4000 cal yr BP and provides insights into how the area has experienced transition from sea to land (Fig. 4). Our reconstruction shows that a group of small islands had uplifted from the sea already before the Litorina Sea transgression around 9000 cal yr BP. By 4000 cal yr BP a significant part of the Pärисpea and Käsmu peninsulas had emerged, however, both still existed as islands and were separated from the mainland by fairly wide and shallow straits. The Lohja basin formed an inner part of Hara Bay, whereas the Käsmu basin was part of a strait. By 3000 cal yr BP a semi-closed lagoon had formed in the area of the present Lohja and Käsmu basins, while the Pärисpea Peninsula had joined with the mainland by a narrow neck and the Käsmu Peninsula by an elongated beach ridge. By 2300 cal yr BP the Pärисpea Peninsula was exposed from the sea and the coastline was quite similar to the modern seacoast; the Lohja basin was isolated from the sea and formed a small coastal lake, whereas the Käsmu basin still existed as a sheltered lagoon with very narrow open connections to the sea to the east and northeast. Lake Käsmu had become isolated and the Käsmu Peninsula had become rather similar to the present-day configuration by 1800 cal yr BP. Previous studies have claimed that Lake Käsmu became isolated from the sea around 800 years ago (Kessel et al. 1986), however, this conclusion is based solely on palaeoshoreline, lithostratigraphical and pollen evidence without any radiocarbon dates.

The diatom evidence described above indicates diatom succession of the Käsmu basin from marine to lacustrine environment and the Lohja basin from lagoonal to limnic environment. The relatively abundant occurrence of pelagic planktonic diatoms with marine/brackish-water affinity in the Käsmu sediment record infers deeper-water conditions and the basin had a semi-opened connection with the offshore sea. The proportion of pelagic diatoms remained high until the basin connection with the sea narrowed; the basin itself turned shallower and littoral periphytic flora became dominant.

The lagoonal phase of both studied sediment sequences is characterized by apparent high abundance of small-sized fragilarioid diatoms with brackish-water affinity. According to Stabell (1985), *Fragilaria* spp. may predominate before, during or after the isolation and for this reason they have not been considered in interpreting the isolation level (Risberg et al. 2005).

A striking feature in the post-isolation diatom stratigraphy from both studied lakes is distinct increase in

freshwater planktonic diatoms that reaches 98% in the Käsmu and 92% in the Lohja record, respectively. Many investigations, which have been based on sediment diatoms, have proceeded from changes in diatom habitat groups (i.e. the ratio of planktonic to periphytic diatoms) as a signal for lake-level oscillations (e.g. Stone & Fritz 2004). Planktonic diatoms contribute frustules to the sediment in deep open-water areas, while periphytic diatoms are primarily associated with shallower littoral habitats closer to shores (Wolin & Duthie 1999). Thus, an increase in the share of planktonic forms is commonly associated with a rise in lake water level (Heinsalu et al. 2008). Alternatively, we assume that post-isolation diatom composition reflects water circulation conditions rather than deeper water depth. Heavily silicified cylindrically shaped and filamentous *Aulacoseira* taxa need turbulence to remain in the water column. Therefore *Aulacoseira* species are common in frequently mixed surface waters, but sink to the lake bottom within long-lasting periods of calm weather. Both isolated lakes were located close to the windy sea coast, thus wind-driven wave motion induced turbulent mixing of the water column and favoured the prevalence of *Aulacoseira*-dominated diatom community.

The shore displacement curve that depicts changes in relative sea level is presented in Fig. 5. In constructing this curve across Ancylus Lake and Litorina Sea high stand, data from the isolation basins and evidences from the raised beaches were considered (Linkrus 1969, 1971; Saarse et al. 2003, 2006, 2010). The reconstructed shoreline displacement curve is relatively regular (Fig. 5) and differs considerably from the previous ones (e.g. Kessel & Raukas 1979). The older curves show up to five transgressive waves during the Litorina Sea and Limnea Sea stages, however, these fluctuations are not radiocarbon dated. Our relative sea-level curve shows regular decrease (Fig. 5) and is in agreement with the curves recently reconstructed around the Gulf of Finland and for the northern part of the Baltic Sea (Seppä et al. 2000; Miettinen 2002, 2004; Lindén et al. 2006; Rosentau et al. 2011; Saarnisto 2012), still differing from the reconstruction carried out in the southern part of the Baltic Sea (e.g. Gelumskaitė 2009). Quite regular changes in sea level over the last 5000 years have been reported from different parts of the globe (Murray-Wallace 2007), including the Baltic coast (Lampe et al. 2011). The diagram shows that water level reached 6 m a.s.l. by 2300 cal yr BP. During this time window the land uplift rate at Lohja was 2.6 mm yr⁻¹. These data are in harmony with the current rate of land uplift (Torim 2004), which is still showing a decreasing trend.

CONCLUSIONS

- The formation of the Pärisepa and Käsmu peninsulas started prior to the Litorina Sea transgression when small islands emerged. By 3000 cal yr BP these islands had joined with the mainland and formed the core of the present-day peninsulas.
- The studied lake basins yielded isolation contacts between 2300 and 1800 cal yr BP when the shoreline reached 6 m a.s.l. at Lohja and 4.5 m a.s.l. at Käsmu. The results of this study overturn the previous conclusion on the isolation of Lake Käsmu approximately 800 years ago.
- A change in sediment lithology accompanied with a change from brackish-water taxa to freshwater diatom assemblage marks the isolation of lakes.
- Before the isolation both basins were shallow-water lagoons.
- Our data confirm a continuous relative fall of sea level in response to glacio-isostatic rebound without distinct sea level oscillations during the last 4000 years.
- The total land uplift rate at 2300 cal yr BP was 2.6 mm yr⁻¹. At present it is 2 mm yr⁻¹.

Acknowledgements. Our sincere thanks are to J. Risberg and L. Gelumbauskaitė for important suggestions and remarks. We are grateful to M. Märss for improving the language. Financial support was provided by the Estonian Research Council (project SF0140021s1s2 and ESF Grant 9031) and DoRa Program.

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Käsmu ja Pärisea poolsaare paleorannajoone rekonstruktsioonid

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Lohja ja Käsmu järvesetete litoloogilise koostise, ränivetikate, magnetilise vastuvõtlikkuse ning AMS ^{14}C dateeringute abil selgitati nende järvede isoleerumine Limneamerest. Nimetatud järved asuvad Põhja-Eestis, vastavalt 5,5 ja 3,9 m üle merepinna. Lohja isoleerus Limneamerest umbes 2300, Käsmu ligikaudu 1800 aastat tagasi. Koostati uuringuala rannasiirde kõver. Rekonstrueeriti Käsmu ja Pärisea poolsaare paleogeograafilised kaardid nelja erineva ajahetke kohta (4000, 3000, 2300 ning 1800 kalibreeritud aastat tagasi), kasutades modelleerimiseks rannamoodustiste andmebaasi ja digitaalset reljeefimudelit.

II Grudzinska, I., Saarse, L., Vassiljev, J., Heinsalu A. 2013. Mid- and late-Holocene shoreline changes along the southern coast of the Gulf of Finland. *Bulletin of the Geological Society of Finland* 85: 19–34.

Mid- and late-Holocene shoreline changes along the southern coast of the Gulf of Finland



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Abstract

In response to glacio-isostatic rebound in Estonia, a relative sea level fall occurred during the mid- and late-Holocene, and as a result, lowland regions in northern Estonia have experienced an evolution from sea to land. The mid- and late-Holocene shoreline changes along the southern coast of the Gulf of Finland were reconstructed, using litho-, bio- and chronostratigraphical proxies from four lakes. The lakes are located within the Gulf of Finland drainage system at different altitudes between 18 and 4 m above the present sea level. The isolation from the sea and the onset of freshwater lacustrine sedimentation occurred in Tänävjärv basin at 5400 cal yr BP, in Klooga basin at 4200 cal yr BP, in Lohja basin at 2200 cal yr BP and in Käsmu basin at 1800 cal yr BP. Through the application of GIS-based analysis, a modern digital terrain model and reconstructed past water level surfaces, we present a series of scenarios of shoreline and palaeogeography changes occurring since 7800 cal yr BP. The land uplift rate, which was approximately 2.8 mm yr⁻¹ 7800 cal yr BP in the surroundings of Tänävjärv, has decreased to 2.2 mm yr⁻¹ at present and that at Lohja from 2.4 to ca 2.0 mm yr⁻¹, respectively. The relative sea level curves show a land uplift decrease, which is nearly linear since the mid-Holocene.

Keywords: lake sediments, stratigraphy, diatoms, absolute age, C-14, paleogeography, sea-level changes, Litorina Sea, Holocene, Estonia

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Editorial handling: Joonas Virtasalo

1. Introduction

Since the deglaciation, the shoreline configuration around the Baltic Sea has been continuous, being controlled by the interaction between local glacio-isostatic recovery and global eustatic sea level rise

(Björck, 1995). In Estonia, the latest research on shoreline displacement has been focused on the earliest stages of the Baltic Sea, i.e. the development of the Baltic Ice Lake (Vassiljev et al., 2011; Vassiljev

& Saarse, 2013), Yoldia Sea (Heinsalu & Veski, 2007), Ancylus Lake (Saarse et al., 1999) and Litorina Sea (Saarse et al., 2009, 2010). However, the shoreline development after the Litorina Sea transgression, which formed a comparatively well developed and therefore easily traceable beach formation on the southern coast of the Gulf of Finland, is considerably poorly studied, many disputable aspects exist and the current data are not always consistent with the results from the neighbouring areas (Miettinen, 2002; Sandgren et al., 2004; Miettinen et al., 2007).

Ramsay (1929) was the first, who described the Litorina Sea beach formations at 53 different locations in Estonia and reconstructed its isobases, which are valid up to the present. Kents (1939) distinguished the Litorina Sea shorelines at five different levels, compiled a shoreline diagram and suggested the uplift gradient 15.5 cm km^{-1} for the Litorina Sea, which has been accepted by many later researches (e.g., Kessel & Raukas, 1967). Detailed studies on the post-Litorina sea level changes are highly limited and the shoreline changes for this time window are poorly constrained. To overcome this problem, we studied four basins, using the isolation method that is widely utilised in sea level reconstructions (e.g., Lindén et al., 2006; Miettinen et al., 2007; Lunkka et al., 2012; Saarnisto, 2012).

The main objectives of the current work are to study mid- and late-Holocene shoreline changes in northern Estonia by providing the chronological control on relative sea level changes. Therefore a series of emerged coastal

lakes, situated at various elevations between the Litorina Sea transgression limit and the present sea level, were investigated. We also aimed at reconstructing the local palaeogeography, compiling a relative sea level curve and testing the existing age-depth models (Saarse et al., 2007; Rosentau et al., 2011), which claims that the sea level has regressed rather evenly over the last 5000 years due to a linear land uplift (Mörner, 1979; Lindén et al., 2006).

2. Site descriptions

The studied lakes Tänävjärv, Klooga, Lohja and Käsmu are located in northern and northwestern Estonia (Fig. 1), on the terraces of the Litorina and Limnea Sea at 18.4 and 3.9 m above the present sea level (a.s.l.), respectively (Table 1). The glacial



Fig. 1. Location of the studied lakes in northern Estonia, showing the positions of the palaeogeographic maps in Figs. 4-6.

Table 1. Morphometric characteristics of the studied lakes.

Site	Latitude N	Longitude E	Area (ha)	Water depth (m)	Altitude (m a.s.l.)	Catchment area (km ²)	Threshold (m a.s.l.)
Lake Tänävjärv	59°10'54''	23°48'42''	138.8	3.0	18.4	4.7	17.4
Lake Klooga	59°18'30''	24°13'57''	131.4	2.5	11.8	5.8	11.8
Lake Lohja	59°32'57''	25°41'23''	56.8	3.7	5.5	12.3	6.0
Lake Käsmu	59°34'56''	25°52'51''	48.5	3.3	3.9	16.5	4.5

deposits covered by marine and aeolian sediments are rather thin, except in buried valleys, where they reach 37–46 m (Kadastik & Ploom, 2000). The relatively flat topography with gentle sloping towards the sea resulted in extensive formation of peat around the investigated lake basins. During the Litorina Sea transgression, all the studied lake basins were inundated by sea water, as the maximum shorelines of the Litorina Sea are recorded at 22.1 m a.s.l. at Tänävjärv, 21.9 m a.s.l. at Klooga, 18.8 m a.s.l. at Lohja and 17.7 m a.s.l. at Käsmu.

Lake Tänävjärv (water level 18.4 m a.s.l., threshold 17.4 m a.s.l.) is an elongated shallow, medium-size semidystrophic lake. The original threshold is located in the southern shore of the lake and is buried under the peat. Prior to AD 1934, Tänävjärv had an outlet to Lake Veskiärv, its area was smaller and water table was lower, (17.9 m a.s.l.; Riikoja, 1934) than at present (Table 1). A ditch between lakes Tänävjärv and Veskiärv has grown over, and consequently, the lake level has been raised about 0.5 m. A set of small beach ridges and dunes has been recognised at 19–21 m a.s.l. in the north and northwest of the lake. The shores of the lake are peaty and partly sandy in the west and east. The catchment is forested by boreal tree species, mostly by *Pinus sylvestris*, and frequently suffers from forest fires (Kangur, 2005).

Lake Klooga (water level and threshold 11.8 m a.s.l.) is located in a north-south oriented depression on the border of the Lahepera-Kloogaranna buried valley (Kadastik & Ploom, 2000) and Pakri klint headland. The lake is surrounded by abraded limestone terraces and a chain of gravelly beach ridges at 22–20 m a.s.l. in the east, dunes in the north and peatland in the west and south (Tammekann, 1940). Lake Klooga is a shallow drainage lake, largely overgrown by emergent aquatic macrophytes. Due to the overabundance of the macrophyte stand, the lake area has decreased to 7 ha over the last 70 years. Small brooks and bottom springs carry calcareous water to the lake and promote precipitation of lacustrine lime. A ditch outflowing to the Vasalemma River is temporarily dry. The western and southern part of the catchment is paludified and forested; the eastern part is covered

by fields and meadows, the northern part by a pine forest. The bottom of the lake is covered by gyttja, calcareous gyttja and silty gyttja with the maximum thickness of 4.5 m in the central part of the basin.

Lake Lohja (water level 5.5 m a.s.l., threshold 6.0 m a.s.l.) is located in a wide Valgejõe-Loobu klint bay which is entirely filled up with sand (Tammekann, 1940). The well-shaped Ancyclus Lake beach ridges have formed an arc 2 km to the south of Lake Lohja, while chains of the Litorina Sea and Limnea Sea beach ridges and dunes, which are parallel to the shoreline, occur in low-lying places. The upper limit of the inner beach ridge lies at 9 m in the west and at 8 m a.s.l. in the north. Lake Lohja is a dark-coloured lake with an outflow brook to the Hara Bay.

Lake Käsmu (water level 3.9 m a.s.l., threshold 4.5 m a.s.l.) is located on the Käsmu Peninsula and is a drainage lake with water that is highly rich in nutrients (Table 1). The lake is bordered by beach ridges at 8 m a.s.l. in the west and a slightly paludified marine plain in the east, covered by different types of boreal forest. The transgressional beach ridges of Ancyclus Lake and the Litorina Sea run 2–3 km to the south of the lake. The water level of Lake Käsmu is regulated.

3. Material and methods

A series of overlapping cores were obtained with a Russian peat sampler from the deepest parts of the basins. 1-m-long core sections were described in field, photographed, sealed in plastic liners and transported to the laboratory and stored in a cold-room. The organic matter (OM) content was quantified by loss-on-ignition (LOI) at 525 °C. The carbonate content was estimated in terms of the difference between LOI at 900 °C and 525 °C multiplied by 1.36. The ignition residue was estimated as mineral matter content. Magnetic susceptibility (MS) was measured with a Bartington MS2E high-resolution scanning sensor from the sediment surface at 1 cm resolution.

Diatom analyses from Lake Tänävjärv (Saarse et al., 1989) and from Lake Käsmu (Kessel et al., 1986) were made by E. Vishnevskaya several years

ago. For this data, diatom taxonomy was modified, diagrams were redrawn and sediment sequences were correlated on the basis of lithostratigraphy and LOI results. For the other sediment records the diatom preparation followed techniques described in Battarbee et al. (2001). Diatom samples were digested in hydrogen peroxide until all OM was removed, hydrochloric acid was added to remove carbonates, and repeated decantation was applied to extract fine and coarse mineral particles. Some drops of the remaining residue were spread over the cover slip, dried overnight and mounted permanently onto microscope slides, using Naphrax medium. At least 400 diatom valves were counted from each subsample under Zeiss Axio Imager A1 microscope at $\times 1000$ magnification and identified to species level. Diatoms were grouped according to their salinity tolerance into brackish/marine, halophilous and freshwater taxa, and according to their habitat into plankton and periphyton. Diatom floras used for the identification and the ecological information were based on the work by Krammer & Lange-Bertalot (1986, 1988, 1991a, 1991b), Witkowski et al. (2000), Snoeijs (1993), Snoeijs & Vilbaste (1994), Snoeijs & Potapova (1995), Snoeijs & Kasperovičienė (1996), Snoeijs & Balashova (1998). Sediment LOI and MS, as well as diatom results were plotted, using the TGVView software (Grimm, 2007).

The radiocarbon dating of macrofossils was performed partly in the Poznan Radiocarbon Laboratory (AMS dates), partly in the Institute of Geology at Tallinn University of Technology (conventional dates). The chronology of the studied sediment sequences is based on the calibration of the radiocarbon dates, using the IntCal09 calibration dataset (Reimer et al., 2009) and the OxCal 4.1 program (Bronk Ramsey, 2009). Radiocarbon dates and lithological data were combined, using the OxCal deposition model (Bronk Ramsey, 2008). In the present study the calibrated ages (cal yr BP) are provided as weighted averages with 2 sigma. Due to the low level of salinity in the Gulf of Finland (Eronen et al., 2001; Miettinen et al., 2007), the reservoir effect in the coastal sediment of Estonia has not been considered. At the same time, results

from an archipelago not far from Stockholm show reservoir ages between 1100 and 400 years (Hedenström & Possnert, 2001), commonly assumed to be in the range of 200 and 400 years (Risberg et al., 2005). The latest studies confirm not only the spatial, but also the temporal difference in the reservoir age since the Litorina Sea (Lougheed et al., 2012).

Palaeogeographical reconstructions are based on GIS analysis in which interpolated water level surfaces were removed from the digital terrain model (DTM; Rosentau et al., 2009). Topographic maps on scales of 1:10 000 and 1:25 000 were used to create a DTM with grid sizes 15 \times 15 m (Tänavjärv, Klooga) and 20 \times 20 m (Lohja, Käsmu). The peat deposits were removed from the DTM, using soil maps on a scale of 1:10 000, whereas the data on peat thickness were obtained from different sources. For the purpose of constructing the relative sea level curve, additional materials from several isolation basins, such as radiocarbon dates and morphometrical data on the raised beaches, were considered (Table 2).

4. Results and discussion

4.1. Environmental conditions

The diatom analysis applied in this study proved to be an effective tool in identifying the position of the isolation contact in the sediment sequence and in defining the related changes in basin salinity and isolation dynamics. The succession of diatom assemblages distinctly records palaeoenvironmental changes induced by the glacio-isostatic uplift and consecutive relative sea level regression through periods of brackish-water environment, isolation from the sea and subsequent lacustrine conditions.

Lake Tänavjärv sediment comprises sand (Tä-1), silt with dispersed OM (Tä-2), silty gyttja (Tä-3) and gyttja (Tä-4; Table 3), and their LOI results are displayed in Figure 2A. MS values are low, even in the silt. Silt with plant remains in core depth between 347 and 342 cm was deposited about 5700–5600 cal yr BP. The basal pre-isolation sediment of the Tänavjärv basin is characterised by

Table 2. Radiocarbon dates considered for the reconstruction of the relative sea level curve.

Site	Basin altitude (m a.s.l.)	Depth (cm)	¹⁴ C date	Calibrated age range, BP (weighted average)	Laboratory ID	Dated material	Reference
Tänavjärv	18.4	316-321	4490±70	4920-5310 (5130±100)	Tln-3306	Gyttja, bulk	Current study
Tänavjärv	18.4	321-324	4600±100	5050-5450 (5270±110)	Tln-3305	Silty gyttja, bulk	Current study
Tänavjärv	18.4	342-347	4930±40	5600-5730 (5660±40)	Poz-42173	Woody pieces	Current study
Klooga	11.8	335-340	3760±40	3990-4230 (4110±60)	Poz-42168	Terrestrial macro remains	Current study
Klooga	11.8	390-395	3840±50	4150-4420 (4280±80)	Poz-42169	Terrestrial macro remains	Current study
Lohja	5.8	365-370	2280±30	2160-2350 (2270±60)	Poz-42171	<i>Pinus</i> bark	Current study
Lohja	5.8	395	2490±35	2440-2730 (2580±80)	Poz-42172	<i>Tilia</i> wood	Current study
Käsmu	3.9	416	1830±30	1700-1860 (1770±40)	Poz-42177	<i>Pinus</i> bark	Current study
Käsmu	3.9	433	1910±30	1790-1930 (1860±30)	Poz-42166	<i>Pinus</i> wood	Current study
Aabla	24	545-555	7250±80	7940-8280 (8080±80)	Tln-3195	Coarse detritus gyttja, bulk	Saarse et al., 2010
Aabla	24	556	7280±50	7990-8190 (8090±50)	Poz-33490	Piece of wood	Saarse et al., 2010
Aabla	24	570-572	6920±40	7670-7840 (7750±50)	Poz-35465	Sand with OM	Saarse et al., 2010
Maarikoja	16.8-17.3		6820±70	7570-7830 (7670±70)	Tln-200	Buried <i>Carex-Phragmites</i> peat	Kessel & Linkrus, 1979
Maarikoja	16.8-17.3		7240±90	7870-8300 (8070±90)	Tln-201	Wood	Kessel & Linkrus, 1979
Vääna	24.7	201.5	6240±40	7070-7270 (7200±50)	Poz-24245	Plant remains	Saarse et al., 2009
Vääna	24.7	220-221	7420±40	8170-8340 (8250±50)	Poz-24267	Plant remains	Saarse et al., 2009
Niitvälja	19.5	280-290	7580±70	8210-8540 (8390±70)	Tln-261	Buried gyttja, bulk	Punning et al., 1980

floristic heterogeneity, mixed occurrence of marine and brackish-water species with relatively frequent freshwater forms (Fig. 3A). The dominant part of the assemblage, however, consists of periphytic saline water tolerant species, while euplanktonic forms are absent. Periphytic brackish/marine diatoms, namely *Diploneis didyma*, *Cocconeis scutellum*, *Hyalodiscus scoticus* and *Campylodiscus chypeus* are the diatoms that are most abundant in the silt layer, accounting for 60–80 % of the assemblage (Fig. 3A). This

diatom evidence is most likely an indication of a sediment accumulation in a shallow-water lagoon, rather than in an open-sea environment. According to radiocarbon dates, the lagoonal phase of the basin lasted for at least 400 years.

The distinct change in the diatom composition from brackish to freshwater species and sediment lithostratigraphy from silty gyttja to gyttja at the core depth of 327 cm indicates the isolation event. The modelled age of the isolation is 5420±130 cal

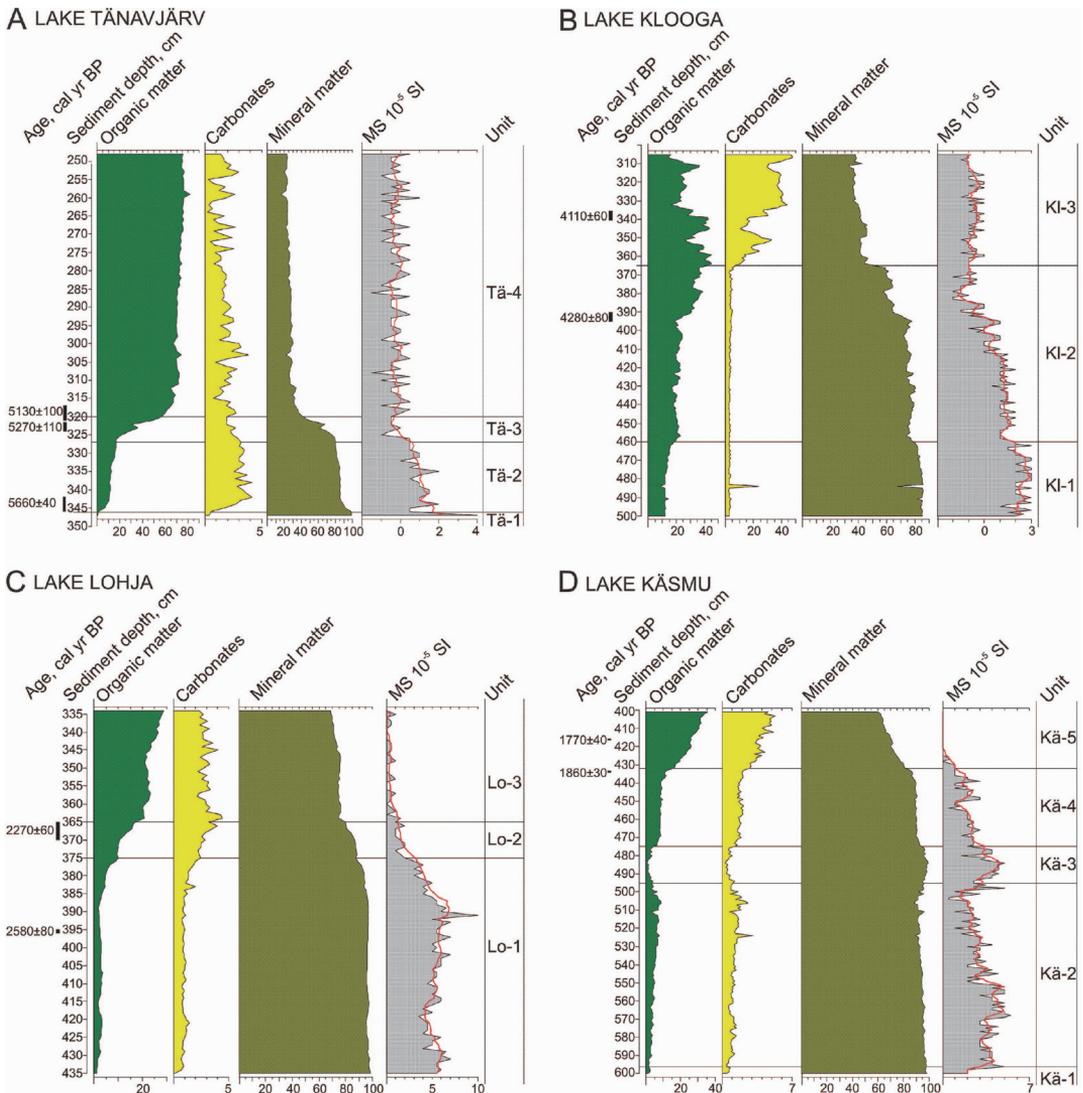


Fig. 2. Sediment organic matter, carbonates and mineral matter content estimated by loss-on-ignition (% of dry weight) and magnetic susceptibility (MS) of Lake Tänavjärv (A), Lake Klooga (B), Lake Lohja (C) and Lake Käsmu (D). The red line in the MS curve is 5-sample moving average.

yr BP (hereafter 5400 cal yr BP). Above the isolation level brackish/marine diatoms disappear, the halophilous taxa decreases and the freshwater taxa, such as *Navicula radiosa*, *Stauroneis phoenicenteron*, *Sellaphora pupula* and *Cymbopleura naviculiformis* increases. The isolation from the sea was probably a rather short-term process.

Initially, the water in the post-isolated basin was rather nutrient rich, confirmed by benthic diatoms such as *Navicula radiosa* and *Sellaphora pupula*. However, the final takeover of diatoms such as *Pinnularia* spp., *Brachysira serians*, *Eunotia* spp. and *Tabellaria fenestrata* implies that the lake underwent a rapid change from alkaline to acidic nutrient-poor

Table 3. Lithostratigraphy of the studied sediment cores.

Site	Depth, cm	Sediment description	Unit
Lake Tänävjärv	245-320	Gyttja, dark brown, slightly consolidated	Tä-4
	320-327	Silty gyttja, dark grey, limit with gyttja sharp	Tä-3
	327-347	Silt with dispersed OM, dark grey	Tä-2
	347-350+	Sand, fine grained, light grey	Tä-1
Lake Klooga	300-365	Calcareous gyttja, beige, subfossil mollusc shells, fluctuating OM	Kl-3
	365-460	Silty gyttja, dark brown, well decomposed OM	Kl-2
	460-500+	Silt with dispersed OM	Kl-1
Lake Lohja	333-364	Gyttja, brown, loose	Lo-3
	364-375	Silty gyttja, greenish grey, at 363 cm richly plant remains	Lo-2
	375-435+	Silt, grey dispersed OM	Lo-1
Lake Käsmu	400-433	Gyttja, dark brown, soft	Kä-5
	433-475	Silt with OM, dark grey	Kä-4
	475-495	Sand, medium size, at lower limit thin OM rich layer	Kä-3
	495-595	Silt with sparse plant remains, dark grey	Kä-2
	595-600+	Sand, fine grained, grey	Kä-1

conditions (Bigler et al., 2000; Dixit & Dickman, 1986) due to paludification of the catchment.

Lake Klooga sediment lithology differs from that of Lake Tänävjärv, containing more OM rich silt (Kl-1) and silty gyttja (Kl-2) that grades into calcareous gyttja (Kl-3; Table 3). The OM content peaks in the upper part of silty gyttja (Kl-2) and lower part of calcareous gyttja (Kl-3; Fig. 2B). The shift in the MS values in general coincides with sediment lithostratigraphical boundaries and is associated with decrease in mineral matter. The calcareous gyttja that terminates the sediment sequence is characterised by low MS values (Fig. 2B). Silty gyttja in the basal part of the sediment sequence (420–365 cm) contains periphytic brackish/marine diatoms, namely *Achnanthes fogedii*, *Campylodiscus clypeus*, *Planolithidium delicatulum* and *Karayevia submarina*, periphytic halophilous taxa, such as *Epithemia turgida*, *E. sorex*, *Hippodonta hungarica* and *Cocconeis placentula*, and small-sized fragilarioid species with brackish water affinity, such as *Pseudostaurosira geocollegarum*, *Pseudostaurosira elliptica* and *Opephora mutabilis* (Fig. 3B). The diatom assemblage suggests brackish water environment, and the AMS ^{14}C dating yielding the age of 4280 ± 80 cal yr BP consequently indicates the period when the basin had been a semi-closed

lagoon with a connection to the open sea.

In the interval between 365–350 cm, cosmopolite salinity-indifferent *Fragilaria* spp. predominate after the disappearance of the brackish-water diatom assemblage. Small-sized fragilarioid taxa are considered to be the pioneer diatoms that have an advantage in a rapidly changing environment (Yu et al., 2004) and thus the peak of *Fragilaria* spp. is regarded as the marker of the transition to the final isolation of the Klooga basin. A dominance of periphytic freshwater diatoms, such as *Cymbella-falsa diluviana*, *Sellaphora vitabunda* and *Achnantheidium minutissimum* from the core-depth of 345 cm indicates the isolation from the sea and suggests a shallow hard-water lake environment. An AMS ^{14}C age of 4110 ± 60 cal yr BP corresponds to the time shortly following the isolation and the modelled age of isolation is 4180 ± 50 cal yr BP (hereafter 4200 cal yr BP).

Lake Lohja lithostratigraphy follows the transition from silt (Lo-1) to silty gyttja (Lo-2) and to gyttja (Lo-3; Table 3). The MS of the sequence is rather stable between the core depths of 435 and 390 cm, decreasing successively between 390 and 360 cm and stabilising in the upper part at low values (Fig. 2C).

The basal part of the core (400–365 cm) shows

high values of benthic brackish/marine diatoms, notably *Planothidium delicatulum* (Fig. 3C). Simultaneously, small-sized fragilarioid epipsammic diatoms with brackish-water affinity, such as *Opephora mutabilis*, *O. guentergrassi* and *Fragilaria gedanensis* are abundant. At the core depth of 360 cm brackish/marine diatoms have declined and planktonic freshwater diatom *Aulacoseira subarctica* predominates together with *A. ambigua* and *A. granulata*, implying a relatively abrupt isolation. An AMS radiocarbon date of 2270 ± 60 cal yr BP, obtained from a pine bark fragment, represents the date that immediately precedes the isolation, and the modelled age of isolation is 2230 ± 70 cal yr BP (hereafter 2200 cal yr BP).

Post-isolation sediments are characterised by the predominance of planktonic freshwater *Aulacoseira* species. The most common diatom *Aulacoseira subarctica* is meroplanktonic and is only present in the water column when there is sufficient turbulence (Gibson et al., 2003), therefore the distinct increase in planktonic diatoms is not related to the lake level rise, but can instead be explained by the location in an open landscape close to a windy sea coast (Grudzinska et al., 2012), where the exposure to wind-induced waves resulted in the turbulent mixing of the water column.

Lake Käsmu lithostratigraphy is rather similar to Lohja site, however, the Käsmu core also includes a sand layer (Kä-3), imbedded into silt (Kä-2, Kä-4; Table 3). The MS shows a wiggly appearance throughout the lithostratigraphical units (Fig. 2D). The diatom composition at the base of the core includes mainly diatoms with marine and brackish-water affinity (Fig. 3D), such as pelagic *Chaetoceros bolsaticus*, *Pauliella taeniata* and *Thalassiosira baltica*, as well as periphytic *Cocconeis scutellum* and *Tabularia fasciculata*, indicating rather open bay-like conditions with brackish-water environment. The diatom-derived isolation at the depth of 400 cm is marked by a decline in brackish-water diatoms and dominance of freshwater planktonic *Aulacoseira* taxa, reflecting the development of a small eutrophic lake. The modelled age of isolation is 1840 ± 30 cal yr BP (hereafter 1800 cal yr BP). The high abundance of planktonic *Cyclotella meneghiniana* at the

biostratigraphic isolation contact is an interesting feature. A peak of *C. meneghiniana* that grows in variable environmental conditions: in brackish waters with elevated nutrient concentration (Weckström & Juggins, 2006), lakes with high conductivity (Saros & Fritz, 2000) or hypereutrophic lakes (Bradshaw et al., 2002) presumably suggests highly elevated water conductivity and nutrient concentrations during the final isolation event.

4.2. Palaeogeography

Based on the results of the current study, several palaeogeographical maps were constructed, which correspond to the following time windows: for Tänävjärv area 6500, 6000 and 5400 cal yr BP; for Klooga area 7800, 4500 and 4000 cal yr BP; for Lohja and Käsmu area 7800, 2200 and 1800 cal yr BP (Figs. 4–6). During the peak of the Litorina Sea transgression about 7800 cal yr BP, numerous beach ridge systems formed along an ancient coastline, stretching continuously in the North Estonian klint bays as a well-developed beach ridge arc at 21 m a.s.l. 15 km SE from Tänävjärv and a ridge system at about 17.5 m a.s.l. 2.5 km to the south from Käsmu. At that time only small islets emerged not far from the studied lake basins that were fully inundated by the sea.

The development and isolation of Lake Tänävjärv was determined by the Audevälja-Harju-Risti-Pedase buried endmoraine ridge and a glaciofluvial delta sediment reaching up to Lake Tänävjärv (Kadastik, 2004). By 6500 cal yr BP the delta plain had partly emerged, becoming subject to wave erosion and wind deflation, and forming a beach ridge/dune landscape to the north and northwest of Lake Tänävjärv (Fig. 4A). Due to the shallow sea, the isolation of Lake Tänävjärv was a rather short-lived process: by 6000 cal yr BP a lagoon with a passage in the south was formed, surrounded by beach ridges in the north and southeast and by a reworked esker ridge or spit in the west and southwest (Fig. 4B). According to the palaeo-reconstruction, Lake Tänävjärv was fully isolated by 5400 cal yr BP (Fig. 4C). These results are consistent with the conclusions made earlier on the basis of

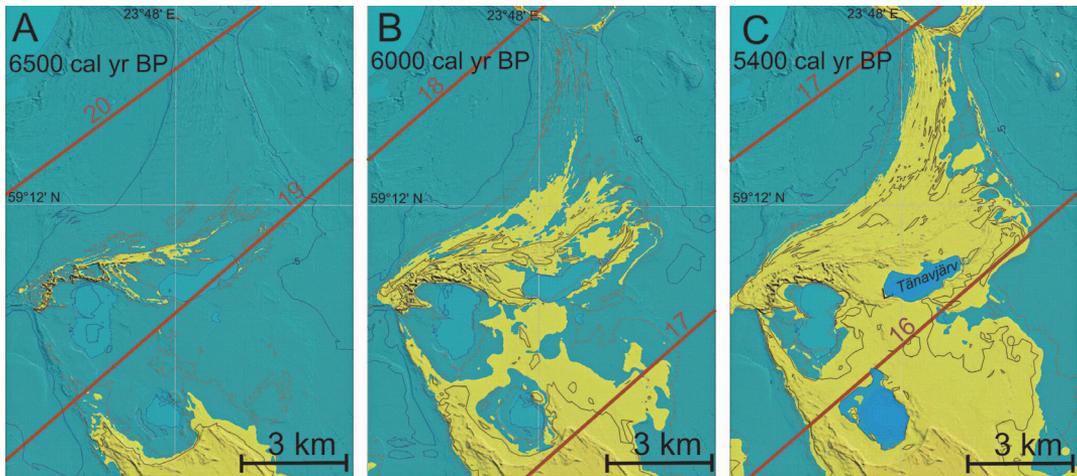


Fig. 4. Palaeogeographic maps of the Tänävjärv area for the time windows of 6500 (A), 6000 (B) and 5400 (C) cal yr BP. Modelled water level surface isobases are indicated by brown lines together with altitudes in meters a.s.l. The shoreline is shown with deviation ± 1 m (black line +1 m, brown line -1 m). Blue line corresponds to the 5 m water depth. The reconstructions are compiled with LIDAR elevation data (Estonian Land Board) to visualize the present day land (yellow) and sea (blue) area.

pollen analyses, according to which sediments corresponding to the isolation were deposited in the second half of the Atlantic period (Saarse et al., 1989).

During the Litorina Sea transgression, a spit was formed in the area to the east of Lake Klooga, isolating a small lagoon. In the surroundings of Niitvälja, a coastal lake was formed (Fig. 5A), where gyttja deposited until 8390 ± 70 cal yr BP (Table 2), later buried by deposits of marine and terrestrial deposits. At 5000 cal yr BP, a beach ridge system developed to the east of the lake, while a tombolo started to form north of the lake, closing the northern connection with the sea by 4000 cal yr BP (Fig. 5C), and only a small passage in the west through the Vasalemma River valley provided the connection with the sea. Obviously, the lake level was slightly higher than the sea level and the passage functioned as a drainage canal of the lake, because the diatom record confirms an isolation around 4200 cal yr BP, which compared to Lake Tänävjärv was tardy due to the depth of the surrounding sea. Therefore, the isolation of Lake Klooga was contingent not only on the land uplift, but also on the development of different beach formations.

At 7800 cal yr BP the coastline of northern Estonia mostly followed the klint escarpment (Fig. 6A) and both Lohja and Käsnu lakes acted as offshore basins with small nearby islets. They maintained their broad connection with the sea until 4000 cal yr BP (Grudzinska et al., 2012). At 3000 cal yr BP, semi-closed lagoons formed in the place of Lohja and Käsnu lakes, with narrow passages between the sea and the lagoons. Lohja and Käsnu lakes became isolated by 2200 cal yr BP and 1800 cal yr BP, respectively, and developed into isolated coastal lakes (Fig. 6). This evidence confirms that Lake Käsnu isolated about 1000 years earlier than previously suggested (Kessel et al., 1986). The isolation history of Lake Lohja in the rump of the klint bay was mainly determined by the post-glacial relief-forming processes, primarily by the beach ridges and dunes. The isolation of Lake Käsnu on the drumlin-like peninsula (Karukäpp, 2004) was also controlled by glacial formations, primarily by a buried esker ridge to the west of the lake. LIDAR maps show that the surface of these drumlin-like forms was jointed by several esker-like ridges orientated to the ice flow direction as can be seen on the Pärisea Peninsula (Fig. 6).

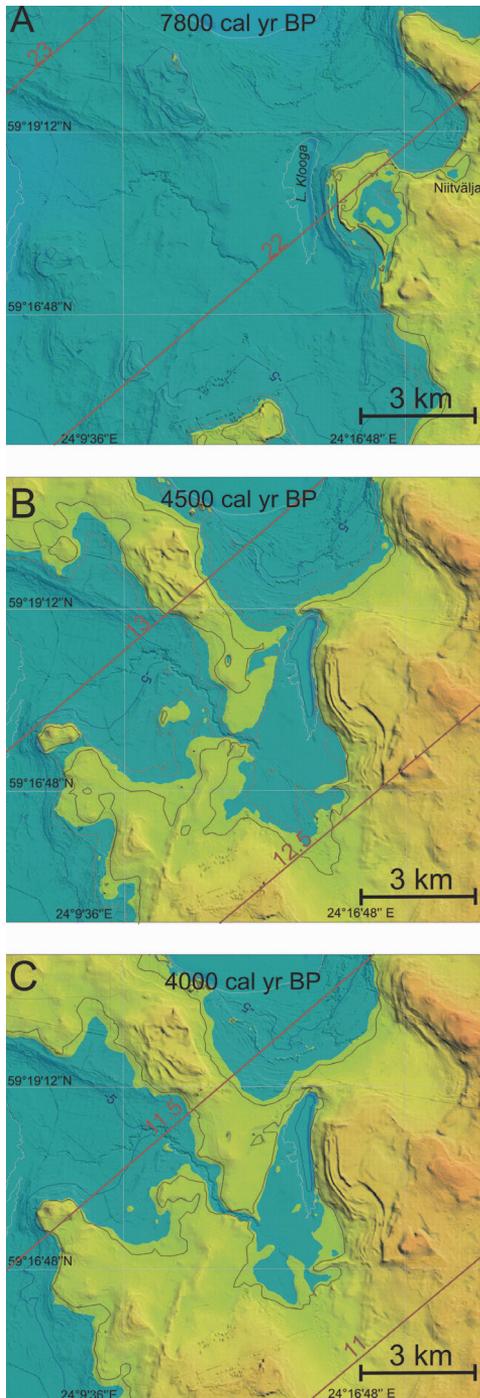


Fig. 5. Palaeogeographic maps of the Klooga area for the time windows of 7800 (A), 4500 (B) and 4000 (C) cal yr BP. For legend explanations see Fig. 4.

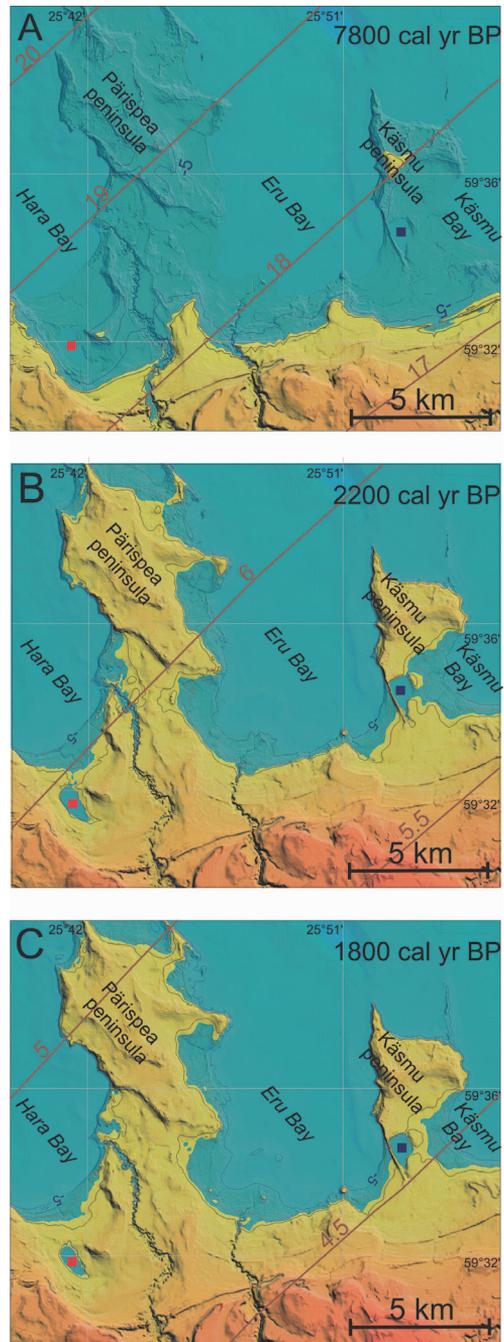


Fig. 6. Palaeogeographic maps of the Lohja and Käsmu area for the time windows of 7800 (A), 2200 (B) and 1800 (C) cal yr BP. For legend explanations see Fig. 4. Lake Lohja is shown by a red square and Lake Käsmu by a blue square.

4.3. Relative sea level curve

Despite several studies conducted in northern and western Estonia, the relative sea level curves that are based on both bio- and chronostratigraphical data have not been constructed until only recently. The relative sea level curves for the Tänävjärv, Klooga, Lohja and Käsmu area were compiled, considering the diatom evidence, lithostratigraphy, ^{14}C dates and geomorphological markers. The presented relative sea level curves are regular from the onset of the Litorina Sea regression (Fig. 7), showing smoothly falling sea level from 22 m a.s.l. down to the present level. The isolation horizon provides regular upwards-younger radiocarbon ages, and biostratigraphic proxies do not show evidence of transgression after 6000 cal yr BP. The smoothly falling relative sea level does not rule out changes in local water level during heavy storms that have been rather frequent over the last century and have caused remarkable damages on the coast (Orviku et al., 2009). In general, the reconstructed sea level curves show a linear trend of sea level lowering that is similar to the reconstructed curves around the Gulf of Botnia (Lindén et al., 2006; Widerlund & Andres-

son, 2011) and the Gulf of Finland (Hyvärinen, 1982; Miettinen, 2002; Miettinen et al., 1999; Saarnisto, 2012), but differs from the previously presented ones concerning the southern coast of the Gulf of Finland (Kents, 1939; Kessel & Raukas, 1979; Raukas & Ratas, 1995) and southern Sweden (Berglund et al., 2005) that displayed several water level fluctuations during the Litorina Sea. Based on recent studies, the Litorina Sea transgression peaked about 7800 cal yr BP (Saarse et al., 2010). The amplitude of the transgression was about 3–4 m, which is compatible with studies carried out in southern Finland in areas with the similar isobase of the land uplift (Miettinen, 2002; Miettinen & Hyvärinen, 1997). Considerably higher magnitude of the Litorina Sea transgression occurred in areas of low uplift, being about 7 m in the surroundings of Pärnu (Fig. 1; Veski et al., 2005). As the Baltic Sea shorelines are widely distributed in northern and western Estonia, they also provide the possibility to examine the pattern of the uplift which they are indicative of (Smith et al., 2000). While the average apparent uplift today in the Tänävjärv-Klooga area is about 2.2 mm yr^{-1} and in Käsmu-Lohja area 2.0 mm yr^{-1} (Torim, 2004), then

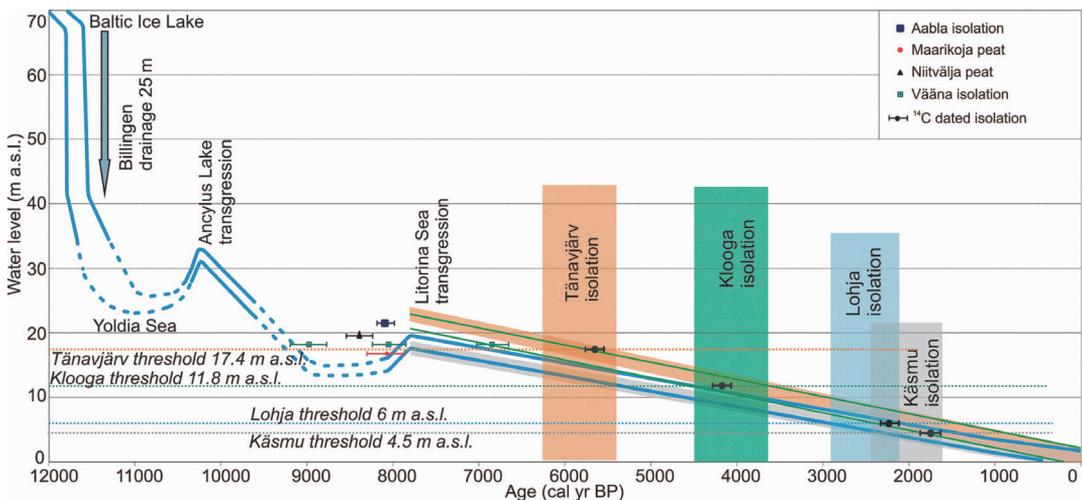


Fig. 7. The relative sea level curves for study areas. Lines mark the modelled possible minimum and maximum water level (green for Lake Klooga, blue for Lake Lohja) and areas mark the modelled possible range of the water level (orange for Lake Tänävjärv and grey for Lake Käsmu). Reconstruction considers errors both in the modelled water levels $\pm 1 \text{ m}$ and in the ages. Boxes in different colours show the isolation age range according to modelled water level. Black circles with error bars indicate the radiocarbon-dated age of the isolation.

according to our calculations the land uplift rates were higher in the mid-Holocene, 2.8 and 2.4 yr⁻¹, respectively, showing a continuous decreasing trend towards the present. If there will not be any rapid rise in the ocean level, the lowering trend of the sea level and apparent uplift in northern Estonia will continue.

5. Conclusions

Threshold elevation and marine limit comparison confirmed that the Litorina Sea water reached 22.1 m a.s.l. at Tānavjärvi, 21.9 m a.s.l. at Klooga, 18.8 m a.s.l. at Lohja and 17.7 m a.s.l. at Käsmu in northern Estonia.

In all four sediment records the succession of diatom assemblages marks distinctly the palaeoenvironmental changes induced by glacio-isostatic uplift and consecutive relative sea level regression through periods of brackish-water environment, isolation from the sea and subsequent lacustrine conditions.

The final isolation contact of the lakes occurred between 5400 and 1800 cal yr BP without notable sea level oscillations during the post-Litorina period.

The isolation of lakes was dependent on the land uplift rate, determined also by the glacial and post-glacial relief forms, such as eskers, limestone terraces, beach ridges and dunes.

The relative sea level curves show a land uplift decrease for the last 6000 years and a smoothly falling sea level.

Acknowledgements

We wish to acknowledge M. Märss for language check. We thank two anonymous reviewers for valuable comments and suggestions. The study was supported by the Estonian Target Financing project SF0140021s12, institutional research funding IUT1-8, ESF Grant 9031 and European Social Fund's Doctoral Studies and Internationalisation Programme DoRa.

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Baltica

BALTICA Volume 27 Number 1 June 2014 : 15–24

doi: 10.5200/baltica.2014.27.02

Biostratigraphy, shoreline changes and origin of the Limnea Sea lagoons in northern Estonia: the case study of Lake Harku

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Grudzinska, I., Saarse, L., Vassiljev, J., Heinsalu, A., 2014. Biostratigraphy, shoreline changes and origin of the Limnea Sea lagoons in northern Estonia: the case study of Lake Harku. *Baltica*, 27 (1), 15–24. Vilnius. EISSN 1648-858X.

Manuscript submitted 15 November 2013 / Accepted 10 April 2014 / Published online 9 June 2014.

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Abstract The paper presents diatom, loss-on-ignition, magnetic susceptibility, and radiocarbon data to reconstruct the depositional history and evolution of Lake Harku, a former Limnea Sea lagoon. Harku is one of the youngest isolated lakes that has been studied bio- and chronostratigraphically in Estonia to date. Based on changes in diatom assemblages, four evolutionary stages in basin development have been recognized (lagoon, semi-enclosed lagoon, transitional and closed lake). Shoreline positions at 2000, 1500, 1000 and 800 cal BP have been reconstructed and displayed on 3D palaeogeographic maps. Lake Harku became isolated from the Limnea Sea at ~800 cal BP, followed by occasional seawater incursions over the next 300 years. Plain landscape, low-lying threshold, and proximity to the sea contributed to extended basin isolation.

Keywords • diatom • radiocarbon dating • loss-on-ignition • 3D palaeogeographic maps • lake isolation

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INTRODUCTION

Beginning in the postglacial period, the Baltic Sea underwent a complicated environmental and geological development (e.g. Berglund 1964; Björck 1995; Hyvärinen *et al.* 1988; Miettinen 2002). In northern regions proximal to the Scandinavian Ice Sheet centre, the sea level regressed, whereas it rose in the southern regions due to differences in glacio-isostatic response. This caused spatial and temporal changes in the Baltic Sea coastline. Regional uplift and apparent sea level lowering in the northern regions resulted in the isolation of coastal water bodies, known as residual/isolated lakes, which emerged during the different stages of the Baltic basin. These isolated water bodies are an excellent sedimentary archive of the evolutionary stages of this coastal region, including the Limnea Sea stage (4500 cal BP up to present). To date, the biostratigraphy and shoreline changes of the Limnea Sea are poorly documented in Estonia. Only two lakes in northern Estonia, which became isolated during the Limnea Sea phase, have been studied bio- and chronostratigraphically (Grudzinska *et al.* 2012).

Lake Harku presents a good setting for reconstructing coastal evolution in this region due to fast sedimentation rate (Saarse 1994), which improve the temporal resolution (Fig. 1). The aim of the current study is to examine environmental changes in Lake Harku, a former Limnea Sea lagoon, focusing on its isolation event, and to present 3D palaeogeographic maps of the shoreline.

GEOLOGICAL SETTING AND SITE DESCRIPTION

The North Estonian klint, one of the most impressive geological monuments in Estonia, is intersected by klint bays and peninsulas. Kakumäe Klint Bay to the west of Tallinn is approximately 10 km long and up to 3 km wide, bordered by Suurupi and Kakumäe klint peninsulas (Fig. 1A, B). A buried valley is incised into the limestone down to 140 m below sea level, reaching the Cambrian bedrock, and is faintly traceable in the topography of the klint bay. This buried valley is filled with till, sand, gravel, varved clay, and sandy-silty deposits of the different stages of the Bal-

tic Sea (Kessel, Pork 1971). After the ice recession, the Baltic Ice Lake flooded the Tallinn area, reaching 110 m above present sea level (Saarse *et al.* 2007) and inundating the entire area of the present city. Because of postglacial isostatic uplift, Tallinn began to emerge from the sea, a trend that is ongoing (Torim 2004).

Lake Harku (59°25'N, 24°37'E) on the western border of Tallinn (Fig. 1A) became isolated rather late in the Holocene and offers a possibility to examine the development and environmental changes in the lake over the past 2000 years. It is a medium size, shallow hypereutrophic lake at 1.2 m above mean sea level (a.s.l.) with an area of $163.3 \times 10^4 \text{ m}^2$

(Tamre 2004), located at the back of the Kakumäe Klint Bay (Fig. 1B). The maximum water depth is only 2.25 m, with an average of 1.7 m. In the 1930s the lake area was smaller – $159 \times 10^4 \text{ m}^2$ and the water level only at 0.9 m a.s.l. (Riikoja 1934). Due to the need to irrigate the surrounding fields, a regulator was installed on its outflow (the Tiskre Brook) in 1974, resulting in a water table rise. Water transparency is less than 1 m and pH varies from 9–10 in the summer to 7.2 in winter. Lake catchment, ca 50 km² in area, is rather densely settled, resulting in an increasing anthropogenic stress. The large residential area of Õismäe is located directly on the south-eastern shore of the lake, with Harku village farther in south, the new residential areas of Harkujärve and Tiskre in the west, and Tabasalu settlement to the northwest (Fig. 1B).

Lake bottom is flat and mostly covered by gyttja, with the maximum thickness 240 cm. Toward the south, the gyttja becomes thinner and wedges out, becoming absent in the littoral zone where only sand is exposed. Harku is a drainage lake, fed by the Harku and Soone Brooks, flowing out to the Kakumäe Bay through the Tiskre Brook (Fig. 1B). Lake shores are flat, partly paludified and covered by meadows, pastures, and a rim of *Alnus* and *Salix* that provide shelter from the wind. In the 1980s, the southern peaty shore was mantled by glaciofluvial sand to create a sandy beach for the Õismäe residents, an action that has been repeated every year.

Aquatic vegetation is scanty and represented by 13 taxa, among which emergent and floating-leaved macrophytes are the most abundant (Mäemets 1977). Since 1950, phytoplankton dominated by the green algae *Scenedesmus quadricauda* started to flourish, causing frequent “water blooms”. To improve the trophic status of the lake, various lake restoration projects have been proposed (Andersen *et al.* 1992), most of which are still awaiting implementation. In 1993–1994, a biomanipulation was carried out for curbing the phytoplankton, but failed to yield the expected results (Leeben *et al.* 2008).

MATERIAL AND METHODS

The bottom deposits of Lake Harku have been examined by several researchers (e.g. Andersen *et al.* 1992; Heinsalu 1993; Saarse 1994; Leeben *et al.* 2008). In winter 2012, series of overlapping cores were obtained with a Russian sampler from the north-western part of the basin (59°25'1.7'' N, 24°36'35.7'' E; Fig. 1B). The uppermost loose sediment was sampled by Willner-type sampler. All 1-m-long core sections were described in field, photographed, sealed in plastic liners, and stored in a laboratory cold-room. Sediment taken by the Willner sampler was cut into 1-cm-thick slices for coming analysis. The lithostratigraphy of the core

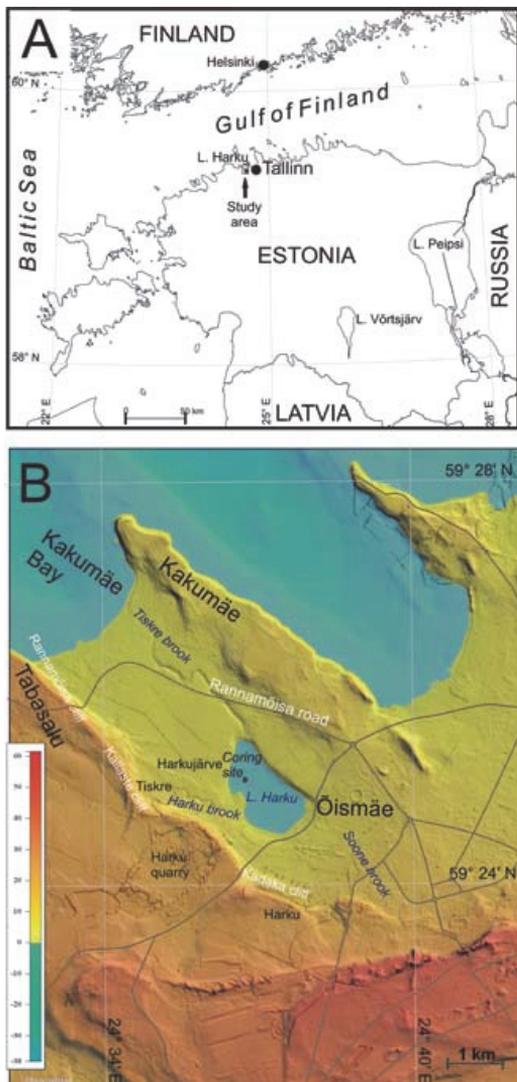


Fig. 1 Location of the study area shown on the overview map (A). Modern topography in the surroundings of Lake Harku with indication of the sampling site (B).

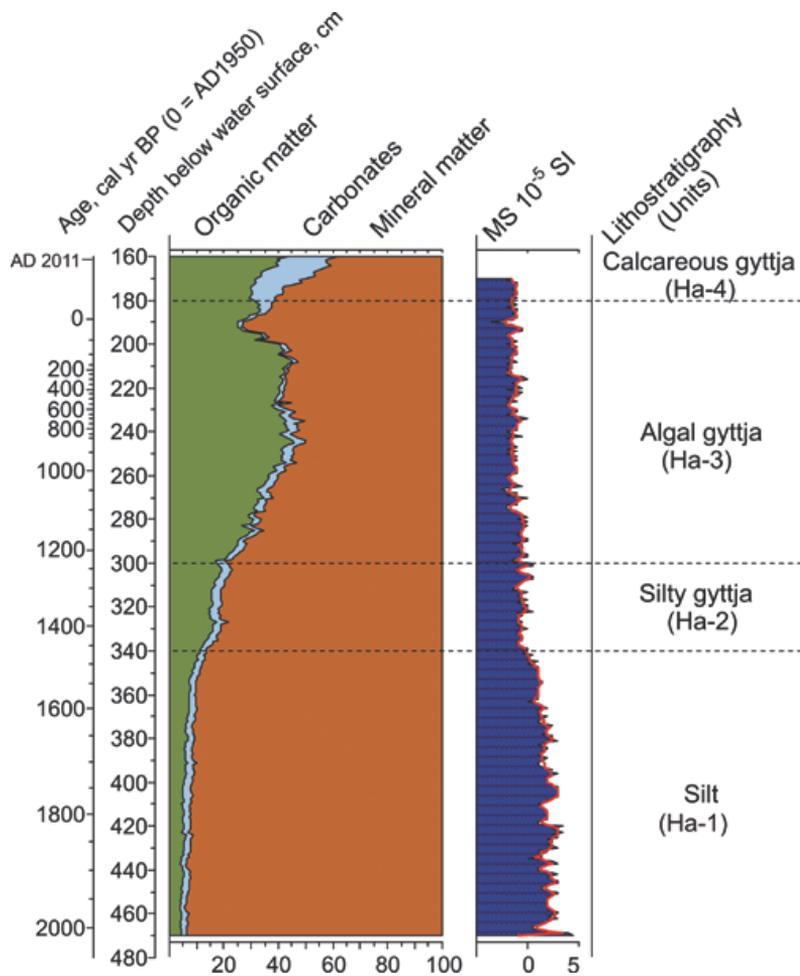


Fig. 2 Loss-on-ignition and magnetic susceptibility from Lake Harku core.

is presented (Fig. 2; Table 1). Continuous 1-cm-thick samples were used for loss-on-ignition (LOI) analysis. The organic matter (OM) was measured at 525°C and expressed in percentages of dry matter. The percentage of carbonates (CaCO_3) was calculated after combustion of LOI residue for 2 hours at 900°C. The mineral fraction was calculated based on the sum of organic and carbonates compounds. Low-field bulk magnetic susceptibility (MS) was measured with a Bartington MS2E high-resolution scanning sensor at 1-cm resolution from cleaned sediment surface covered with a thin plastic film (Fig. 2).

The diatom samples were prepared by following techniques described in Battarbee *et al.* (2001). Sediment samples were digested in hydrogen peroxide to remove all OM, hydrochloric acid was added to remove CaCO_3 , and repeated decantation was applied to extract fine and coarse mineral particles.

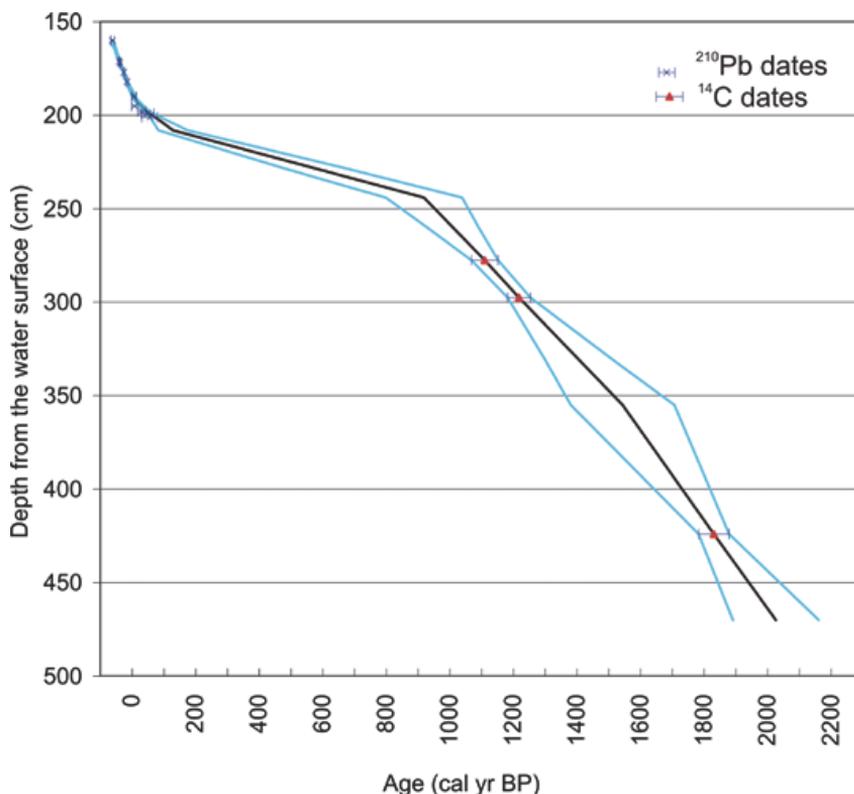
Table 1 Lithology of Lake Harku core

Depth, cm	Lithology
0–160	water
160–170	calcareous gyttja, dark greenish grey, loose
170–180	calcareous gyttja, greyish green, loose
180–240	algal gyttja, greenish brown
180–300	algal gyttja, dark brown
300–340	silty gyttja, greyish
340–470+	silt, dark grey

Some drops of the remaining residue were spread over the cover slip, dried overnight and mounted permanently onto microscope slides with Naphrax medium. Between 500 and 600 valves were counted from each sub-sample under Zeiss Axio Imager A1 microscope at $\times 1000$ magnification and identified to species level in order to estimate the percentage abundance of taxa.

Table 2 Radiocarbon measurements of Lake Harku sediments

Depth, cm	Age, ¹⁴ C BP	Calibrated age, cal BP (average)	Lab. No	Material
275–280	1185±30	1065–1170 (1120±55)	Poz-51453	Plant remains
295–300	1265±30	1175–1260 (1220±45)	Poz-51454	Plant remains
424	1895±35	1750–1890 1820±70	Poz-49185	Wood

**Fig. 3** Age-depth model considering lithological boundaries with respect to loss-on-ignition (see Fig. 2). The black line is weighted average of radiocarbon dates with error bars (blue lines) at one-sigma.

The 170–350 cm core interval was used for diatom analysis. Diatoms were grouped according to their salinity tolerance into marine/brackish, halophilous, small-sized fragilarioid taxa with brackish water affinity, small-sized fragilarioid taxa preferring fresh water, indifferent, freshwater, and unidentified taxa. Habitat classification included planktonic, small-sized fragilarioid, and periphytic taxa. Diatom floras used for the identification and ecological information were based on well-established sources (Krammer, Lange-Bertalot 1986, 1988, 1991a, b; Snoeijs 1993; Snoeijs, Vilbaste 1994; Snoeijs, Potapova 1995; Snoeijs, Kasperovičienė 1996; Snoeijs, Balashova 1998; Witkowski *et al.* 2000).

Macrofossils for radiocarbon dating were extracted by soaking 5-cm-thick samples (with a volume of ~250 cm³) in a solution of water and Na₄P₂O₇. After sieving through a 0.20 mm mesh, the material was

dried at 70°C. Unfortunately, plant macrofossils were very rare as gyttja largely composed of algae, which was confirmed from measurements of N, C, H concentrations and the C/N ratio (Saarse 1994). An age-depth model was produced based on ²¹⁰Pb measurements (Leben *et al.* 2008) and AMS radiocarbon dates of macrofossils (Poznan Radiocarbon Laboratory) (Fig. 3; Table 2). Radiocarbon dates were calibrated at one-sigma confidence level using the IntCal09 calibration dataset (Reimer *et al.* 2009) and the OxCal 4.1 program (Bronk Ramsey 2009) and were combined with lithological data according to the OxCal deposition model (Bronk Ramsey 2008). The present study applied calibrated ages as weighted averages before present (cal BP, 0=AD 1950) (Table 2). ²¹⁰Pb dates permitted extending the chronology up to AD 2011. Diatom, LOI, and MS results were plotted, using the Tilia v.1.7.16 software (Grimm 2011).

Palaeogeographic maps are based on a GIS analysis in which interpolated water level surfaces were removed from the digital terrain model (DTM; Rosentau *et al.* 2009). Topographic maps at scales of 1:2000; 1:10 000, and 1:25 000 were used to create a DTM with grid size of 10×10 m. Palaeogeographical maps for different time windows were compiled based on the assumption that a decrease in the land uplift after the Litorina Sea transgression occurred linearly (Mörner 1979; Yu *et al.* 2007; Rosentau *et al.* 2012) and that global sea level remained nearly constant (Lambeck, Chappell 2001).

RESULTS

Lithostratigraphy and chronology

A lithostratigraphic transect along the lake reveals that the thickness and composition of sediments in Lake Harku is quite similar (Saarse 1994). A comparison of the LOI results of earlier studied cores (Saarse 1994) with the master cores obtained in 2005 and 2012 shows relatively good consistency. The sediment composition changes gradually from silt (Ha-1) to silty gyttja (Ha-2), algal gyttja (Ha-3) and calcareous gyttja (Ha-4; Fig. 2). Silt (core depth of 470–340 cm) contains about 5% OM, in silty gyttja (340–300 cm) the level of OM increases to 20% and in algal gyttja to 40% (Fig. 2). According to the age-depth model (Fig. 3), the silt was deposited between 2030 and 1480 cal BP, silty gyttja between 1480 and 1230 cal BP, algal gyttja from 1230 cal BP to AD 1956, and calcareous gyttja between AD 1956 and 2011. The maximum OM values measured along the 250–200 cm core interval correspond to 950–70 cal BP. The low content of CaCO₃ is typical of other fore-klint lakes (Saarse 1994), except for the topmost gyttja where it rapidly increased to 20%, due to the liming of soils in the lake catchment and by the estab-

lishment of the Harku quarry in 1954 (R. Voog pers. comm.). As Harku sediment are composed of organic, quartz-rich and carbonate components which are diamagnetic, MS values are low. Bulk MS gradually decreased upsection, from 2–3×10⁻⁵ SI units between 470–380 cm, decreasing to zero between 380–335 cm, and remaining negative upsection (Fig. 2).

Diatom stratigraphy

A total of 117 diatom taxa representing 51 genera were identified in the 170–350 cm core interval. The most common diatom species are displayed (Fig. 4). The diatom assemblage in the basal part of the core (350–295 cm) is dominated by small fragilarioid taxa with brackish-water affinity (35–50%), primarily represented by *Pseudostaurosira subsalina*, *Opephora mutabilis*, and *Fragilaria sopotensis*. Marine/brackish-water species are represented by the planktonic *Chaetoceros muelleri* var. *subsalsum*, the periphytic *Planothidium delicatulum* and *Navicula peregrina*, whereas the planktonic halophilous *Cyclotella meneghiniana* occurs at 330–310 cm, indicating temporary nutrient enrichment in the embayment (Weckström, Juggins 2006) (Fig. 4).

Around the depth of 295 cm, *Martyana schulzii* and *Opephora guenter-grassi* disappeared, the content of *Pseudostaurosira subsalina* and *Opephora mutabilis* decreased, and that of the *Pseudostaurosira geocollegarum* and *Opephora krumbeinii* increased substantially. Shortly before the marked increase in the freshwater taxon *Stephanodiscus parvus*, the marine/brackish planktonic *Chaetoceros muelleri* var. *subsalsum* reaches its maximum value of 7%.

In the transition zone (265–225 cm), the most significant feature of the diatom flora is the sharp increase in planktonic freshwater taxon *Stephanodiscus parvus* (up to 32%) and halophilous *Cyclotella meneghiniana* (up to 11%), indicating nutrient enrichment (Ander-

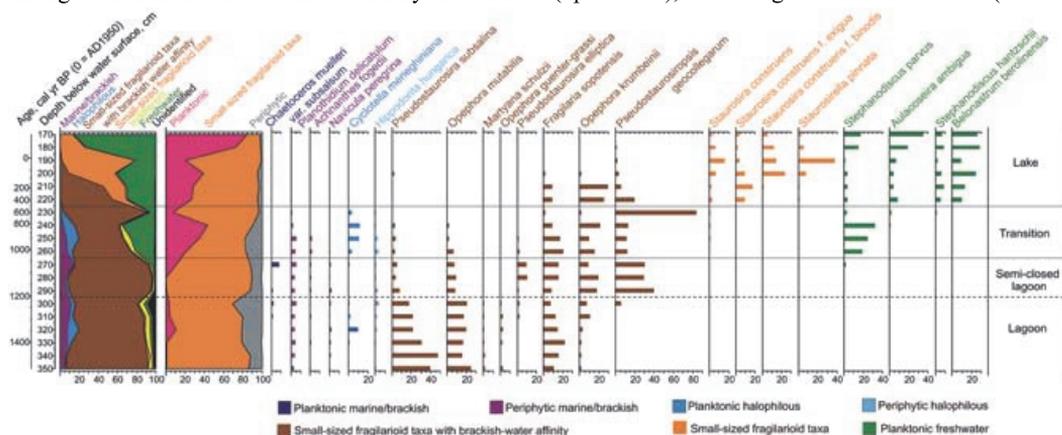


Fig. 4 Percentage diagram of selected diatom taxa from Lake Harku.

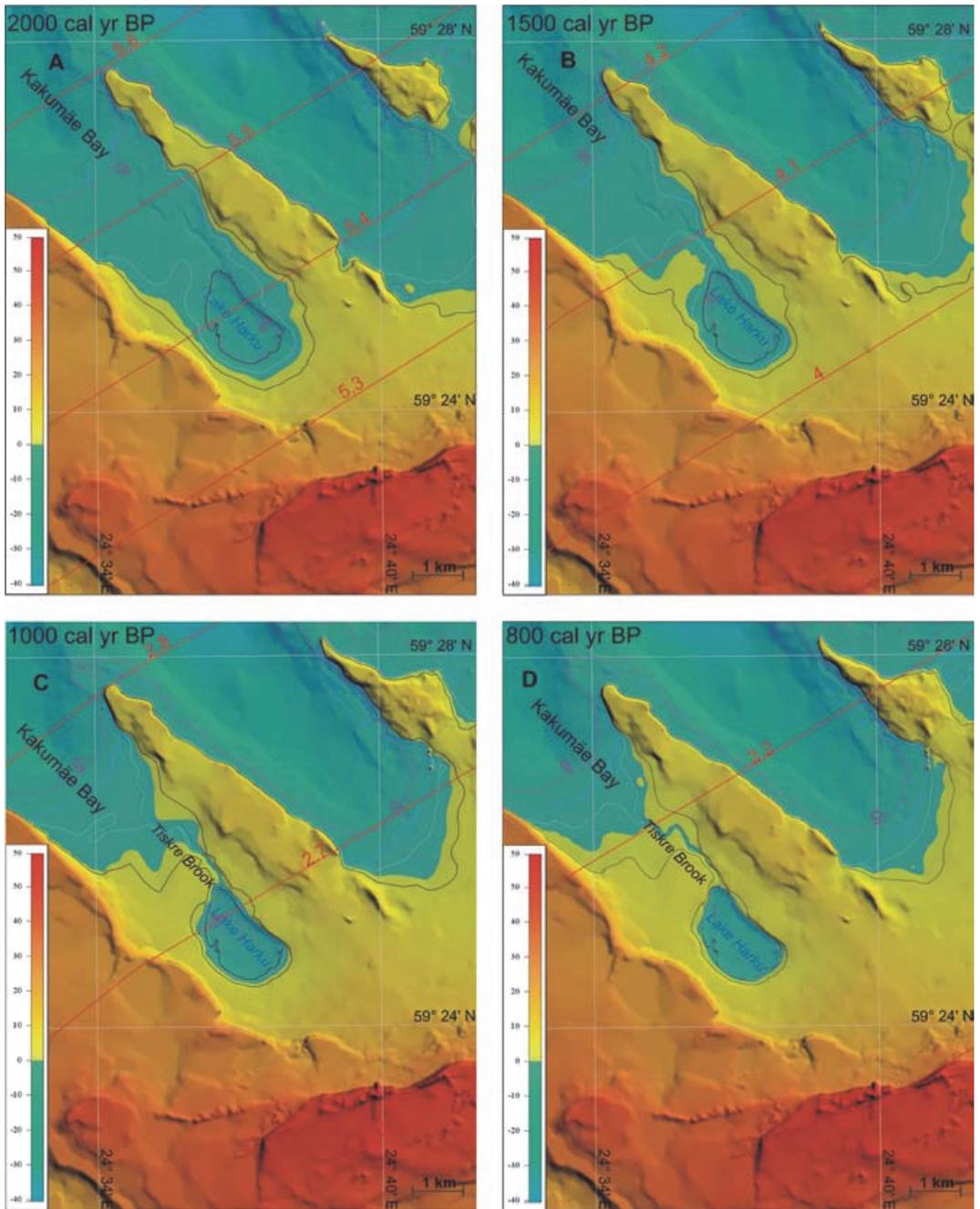


Fig. 5 Palaeogeographic maps of 2000 (A), 1500 (B), 1000 (C) and 800 cal yr BP (D) time windows. Red lines are the water level surface isobases.

son 1990; Witak 2013). These taxa are accompanied by small fragilarioid species with brackish-water affinity, such as *Pseudostaurosiraopsis geocollegarum*, *Opephora krumbeinii* and *Fragilaria sopotensis*, and the periphytic marine/brackish-water *Planolithidium*

delicatulum. Disappearance of *Achnanthes fogedii*, which is a typical species of the Litorina Sea (Snoeijs, Kasperovičienė 1996; Witkowski *et al.* 2000), and the halophilous *Hippodonta hungarica* at the depth of 240 cm indicates reduced influx of brackish water

into the basin. Just before the disappearance of marine/brackish and halophilous taxa at 230 cm, there is a sharp decrease of *Stephanodiscus parvus* from 32 to 2% and an increase of *Pseudostaurosiraopsis geocollegarum* from 13 to 84%.

A distinct increase in the abundance of diatom taxa that prefer freshwater conditions was observed in the uppermost part of the sediment sequence between 225 and 170 cm. Planktonic freshwater taxa are represented by *Belonastrum berlinensis*, *Stephanodiscus hantzschii*, *Aulacoseira ambigua* and *Stephanodiscus parvus*, whereas the main components of the small fragilarioid taxa are *Staurosira construens* f. *exigua*, *Staurosira construens* f. *binodis*, *Staurosirella pinata* and *Staurosira construens*.

DISCUSSION

The application of multiple palaeoenvironmental indicators, such as the diatom analysis, LOI, MS, and establishment of lead and AMS-radiocarbon chronology, aided in constraining the timing of basin isolation and reconstructing the palaeoshoreline during the past 2000 years. To exhibit the spatial and temporal shoreline changes, 3D palaeogeographic maps were constructed for the time windows of 2000, 1500, 1000 and 800 cal BP (Fig. 5). The 2000 cal BP situation shows the position of the shoreline during the phase when the palaeo-Harku basin was located at the head of the Kakumäe Klint Bay. The lake depression and northwestern part of its catchment were entirely covered by the seawater (Fig. 5A). By 1500 cal BP, Harku had transformed into a lagoon, connected with the sea via a 300–400-m-wide pass through the present valley of the Tiskre Brook (Fig. 5B). This wide connection maintained brackish conditions in the Harku basin and domination of the marine/brackish diatom assemblage (Fig. 4). Due to land uplift, the connection with the sea continuously narrowed, resulting in a semi-enclosed lagoon and accumulation of OM rich gyttja with very low MS values (Fig. 2). Based on the age-depth model, the semi-enclosed lagoon phase lasted for approximately 160 years. According to a simulation, the passage between the sea and the lagoon was still rather wide, which ensured brackish conditions in Harku basin supported by the dominance of small-sized fragilarioid taxa with brackish water affinity (Fig. 4). By 1000 cal BP, the passage to the sea through the Tiskre valley still existed, narrowing further to 80–100 m (Fig. 5C).

Gradual changes in the diatom assemblages from marine/brackish and halophilous to freshwater indicate the evolution of the basin from a lagoon to a semi-enclosed lagoon, a transition phase (a closing basin with intermittent brackish-water influx) and finally a freshwater lake no longer affected by marine water incursion.

The co-appearance of marine-brackish (e.g. *Planolithidium delicatulum*, *Achnanthes fogedii*), halophilous (e.g. *Cyclotella meneghiniana*, *Hippodonta hungarica*) and freshwater (*Stephanodiscus parvus*) diatoms in the transition phase at the core depth of 265–225 cm, (roughly between 1000 and 500 cal BP) demonstrates that the isolation of Harku was a long-lasting event. During this phase, the nutrient load in the lake increased, which can be concluded from diatoms *Stephanodiscus parvus* and *Cyclotella meneghiniana*. Similar changes in the trophic state have been observed in many isolation basins (e.g. Grudzinska *et al.* 2012; Seppä, Tikkanen 1998; Westman, Hedenström *et al.* 2002; Yu *et al.* 2004). The enhanced nutrient content in Lake Harku could be explained by two factors: 1) occasional mixing of brackish and fresh water that promotes biological productivity and enrichment with organic compounds (Head 1976, cited after Bechtel *et al.* 2007), and 2) an intense nutrient input from the sparsely vegetated catchment area (Seppä *et al.* 2000). Long-term influxes of brackish water could be explained by the proximity of the sea to Lake Harku, by the wide and low-lying threshold (~2.5 m a.s.l.) and flat topography. This facilitated seawater inflow during heavy storms, although its infiltration through coastal sand cannot be ruled out. Such circumstances made it difficult to determine the exact age and level of the isolation.

One of the indicators of isolation is the mass occurrence of *Fragilaria* spp. (Seppä *et al.* 2000); however, according to Stabell (1985), the peak in *Fragilaria* spp. could occur before, during or after the isolation of the basin from the sea. In case of Harku, *Fragilaria* spp. are the dominant diatoms down to the core depth of 190 cm (Fig. 4). In order to recognize the transition from the marine-brackish to freshwater environment, small fragilarioid taxa with brackish-water affinity are separated from *Fragilaria* spp. that prefers freshwater conditions. The occurrence of *Fragilaria* spp. with brackish-water affinity showed a sharp decrease at 225 cm (ca 500 cal BP), accompanied by the appearance of several freshwater diatom species (Fig. 4), thereby indicating that the lake ecosystem was no longer affected by saline water.

According to this scenario, the Harku basin became isolated from the sea prior to 800 cal BP (Fig. 5D), which is in disagreement with the diatom assemblages that indicate that the brackish-water conditions lasted longer. This discrepancy has obviously resulted from marine water influxes and other aforementioned factors. Previous studies have claimed that the Harku basin became isolated from the sea considerably earlier, approximately 1500 years ago (Saarse 1994); however, this estimation has been solely based on pollen,

lithostratigraphic, and morphologic evidence, in the absence of radiocarbon dates.

During the last 800 years, the shoreline has prograded 2 km in the northeast and 3 km toward northwest and the land has uplifted ca 250 cm with the rate of 0.31 mm per year, which is similar to the value suggested by Künnapuu (1970). However, some scenarios argue that the predicted continuing relative sea-level fall could turn into a relative sea-level rise, even in Estonia (Rosentau *et al.* 2012), based on a considerable acceleration of global sea level rise in the 22nd century (Jevrejeva *et al.* 2012). An average Baltic Sea level rise was calculated 1.4 ± 0.4 mm yr⁻¹ for the 20th century (Rosentau *et al.* 2012), consistent with the global sea-level rise 1.48 ± 0.26 mm yr⁻¹ between 2003 and 2010 (Jacob *et al.* 2012). In contrast, N.-A. Mörner (2004) predicted that by AD 2100, the Baltic Sea level change will be only $+10 \pm 10$ cm and there will be no fear of massive flooding.

CONCLUSIONS

The development of Lake Harku during the past 2000 years has been a rather complex process, exhibiting a distinct marine phase. Based on diatom assemblages, four evolutionary stages (lagoon, semi-enclosed lagoon, transitional lake, and closed lake) have been identified. The isolation of Harku basin was a long-lasting process due to flat topography, wide and low-lying threshold and proximity to the sea, which promoted the incursion of marine water during heavy storms. According to palaeoenvironmental simulations, the basin became isolated from the Limnea Sea shortly before 800 cal BP, but was influenced by seawater for at least 300 more years, which favoured the survival of brackish water diatoms in the already isolated basin.

Minerogenic sedimentation in Harku lasted until ca 1500 cal BP, followed by the deposition of silty gyttja up to ca 1250 cal BP, and culminating in algal gyttja, which became more calcareous over the past 60 years. A sharp increase in calcareous compounds in the upper part of the sequence is attributed to the liming of soils, establishment of the quarry in the lake catchment, and water pollution through rising bioproductivity.

Acknowledgments

The referees Professor Anto Raukas (Tallinn) and Dr. Vaida Šeirienė (Vilnius) are acknowledged for critical remarks and suggestions. Professor Ilya V. Buynevich (Philadelphia) is thanked for constructive comments on the final manuscript. The study was supported by the institutional research funding IUT 1-8, ESF Grant 9031 and Doctoral Studies and Internationalisation Programme DoRa.

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Relative sea level changes and development of the Hiiumaa Island, Estonia, during the Holocene

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Vassiljev, J., Saarse, L., Grudzinska, I., Heinsalu, A., 2015. Relative sea level changes and development of the Hiiumaa Island, Estonia, during the Holocene. *Geological Quarterly*, 59 (3): 517–530, doi: 10.7306/gq.1227



Three sediment cores (Loopsoo, Tihu, Prassi) from Hiiumaa Island (Estonia) were investigated using diatoms, lithological proxies, magnetic susceptibility, geochronological dates and incorporated with the previously studied Kõivasoo site, aiming to reconstruct the development of the island and shoreline changes during the Litorina Sea and the Limnea Sea. The highest level of the Litorina Sea shoreline near Kõivasoo is at 27.6 m a.s.l., and it occurred during the Initial Litorina Sea. Within the Litorina Sea transgression, 7800 cal yr BP, relative sea level reached 24.9 m a.s.l. at Kõivasoo, 24.1 m a.s.l. at Loopsoo, 23.6 m a.s.l. at Tihu, and 21.5 m a.s.l. at Prassi. Kõivasoo became isolated from the sea about 8500 cal yr BP, Loopsoo between 7100 and 6800 cal yr BP, Tihu around 4800 cal yr BP, and Prassi about 2500 cal yr BP. Presently gained data from Hiiumaa Island confirm that the Litorina Sea regressed consistently during the last 8000 years due to progressively declining isostatic rebound. The present study is also illustrated by 3-dimensional palaeogeographic maps of the Hiiumaa Island development.

Key words: Litorina Sea, Limnea Sea, lithology, diatoms, relative sea level changes, Estonia.

INTRODUCTION

Large islands in the Baltic Sea Basin (BSB) have long time attracted interest of researchers as they show marginality in geological and vegetation history and climate conditions (e.g., Luha et al., 1934; Königsson, 1968; Sepp, 1974; Svensson, 1989). Hiiumaa Island experienced significant coastline changes during the Holocene due to interplay between sea level change and isostatic land uplift (Kessel and Raukas, 1967, 1979; Sepp, 1974; Saarse, 1994; Raukas and Ratas, 1996; Hang and Kokovkin, 1999; Saarse et al., 2003). The Ancylus Lake and the Litorina Sea, stages of the BSB, left behind numerous beach ridges, wide spectrum of scarps, and several ancient lagoons, which are potential sites to study water level changes, island development and early human colonization (Kriiska and Lõugas, 1999). Coastal formations of the Ancylus Lake are now positioned up to 45 m a.s.l., that of the Litorina Sea up to 27.6 m, and of the Limnea Sea up to 12.8 m a.s.l. (Kents, 1939), proceeded from the postglacial land uplift that is nowadays approximately 2.5 mm yr^{-1} on Hiiumaa (Torim, 2004).

The current study focuses on the development of the Litorina Sea whose onset is marked with the establishment of the connection between the BSB and the ocean around

9800 cal yr BP (Andrén et al., 2000; Berglund et al., 2005), since when the water levels in the BSB and the ocean were in equilibrium. The Mastogloia Sea has been recognized between the freshwater Ancylus Lake and the brackish-water Litorina Sea as a transitional diatom-stratigraphic unit (Kessel and Pork, 1974; Cker et al., 1988; Hyvärinen et al., 1988, 1992; Haila et al., 1991) based on the presence of weakly brackish-water diatom assemblages particularly in the littoral sequences (Hyvärinen, 1984, 2000), whereas in the offshore sequences such a transitional unit is commonly absent (Ignatius et al., 1981). Lately, this sub-stage was renamed Early (Initial) Litorina Sea, marking penetration of saline water to the BSB about 9800–8500 cal yr BP (Andrén et al., 2000; Berglund et al., 2005; Harff et al., 2011). The slightly brackish-water diatoms occur in the sediment sequences of Finland since 8800 cal yr BP (Eronen, 1974), and about 8500 cal yr BP in western Estonia (Hyvärinen et al., 1988). However, the typical brackish-water mollusc fauna migrated later, when the salinity of the sea had reached 15–20‰ (Hyvärinen et al., 1988). The transgression peak of the Litorina Sea is time-transgressive occurring later in the southern parts of the BSB, i.e. in areas with a slower land uplift rate (Miettinen and Hyvärinen, 1997; Hyvärinen, 2000; Saarse et al., 2000; Miettinen, 2002, 2004).

The current paper summarizes the main results of lithostratigraphical and diatom biostratigraphical analysis, and ¹⁴C radiocarbon dating applied to studies of shore displacement on the Island of Hiiumaa. These data from four isolated basins is then used in GIS-analysis with aim to reconstruct the development of Hiiumaa Island and to illustrate its 3-dimensional (3D) palaeogeographic maps.

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Received: September 24, 2014; accepted: March 19, 2015; first published online: April 9, 2015

MATERIAL AND METHODS

STUDY AREA AND SITE DESCRIPTION

In the current study, three basins: Tihu, Loopsoo and Prassi (Fig. 1) were examined and incorporated with the earlier study by Königsson et al. (1998) and Saarse et al. (2000) from Kõivasoo.

Kõivasoo is a small raised bog, an ancient Litorina Sea lagoon on the Kõpu Peninsula at 27.5 m a.s.l. (Fig. 1B), surrounded by wide spectrum of sandy beach formations. It has a narrow threshold in the south at an elevation of 27 m a.s.l., which was deepened by ditching the bog. Pollen and diatom records of Kõivasoo deposits, taken from the central part (58°54'32"N; 22°11'56"E), were published earlier (Sarv, 1981; Sarv et al., 1982; Königsson et al., 1998; Saarse et al., 2000).

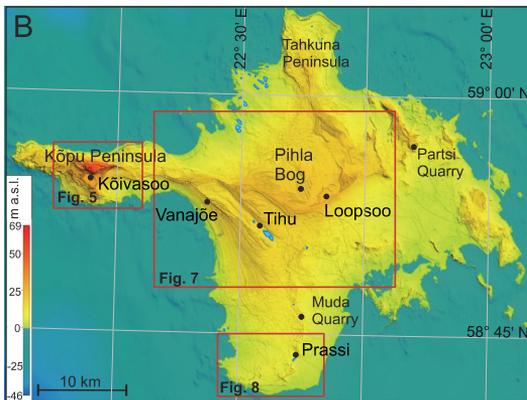
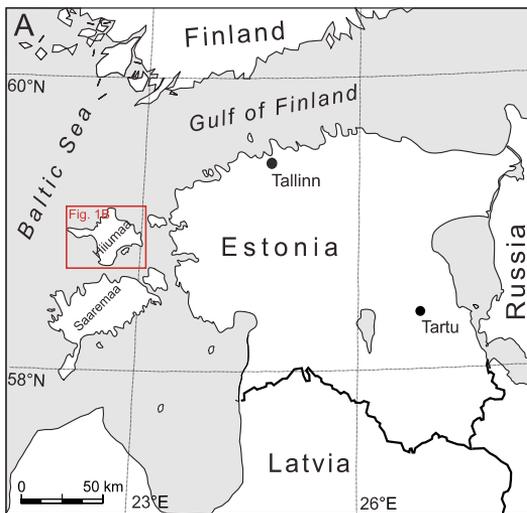


Fig. 1. Overview map of the study area (A) and digital terrain model (DTM) of Hiiumaa Island (B)

Black dots mark the study sites and red squares the areas of palaeogeographical reconstruction

Archaeological remains found on the beach ridges east to Kõivasoo reveal a seasonal settlement during the Late Mesolithic and Early Neolithic (Kriiska and Lõugas, 1999). Typical Litorina Sea mollusc fauna has been identified near the settlement site at 26 m a.s.l. (Moora and Lõugas, 1995). More details about the pollen stratigraphy and radiocarbon dates are given in Königsson et al. (1998), and the development of the Kõpu peninsula is discussed by Saarse et al. (2000).

Loopsoo Bog is located in the central part of Hiiumaa (Fig. 1B), close to the Ordovician and Silurian bedrock boundary (Eltermann, 1993a). The present bog surface lies at 21.5–22 m a.s.l., has an area of 128 ha and is surrounded by a bow-shaped beach ridge. The presently observed threshold at 21 m a.s.l. at the outflow ditch in the eastern part of the bog was probably ca. 1 m lower and was covered by sand during isolation. Samples for the current study were taken from the southern part of the bog (58°53'41"N; 22°40'21"E). Thickness, distribution and properties of the peat were studied by Orru (1995), but the peat layers were not dated. Loopsoo is surrounded by a thin woody rim and field patches. The nearest dated site is Pihla Bog (Fig. 1B), 2 km north-west of Loopsoo, where the basal peat at the contact with sand at about 14 m a.s.l. has been dated to 3530 ± 230 cal yr BP (Liiva et al., 1966).

Lake Tihu Keskjärv (hereafter Tihu) in the western part of Hiiumaa is a small, elongated, shallow (maximum water depth 0.6 m) overgrowing (2.7 ha) semidystrophic lake in a paludified valley-like depression at 14.5 m a.s.l. (Fig. 1B). It is flanked by ridges and dunes reaching up to 27 m a.s.l., and shoaling towards the south-east. The base of the nearest beach ridge was levelled at 15.5 m a.s.l. (Ratas, 1976), and a threshold at 14.4 m a.s.l. is located in the southeastern part of the valley. Here, brownish till with erratic clasts and marine sands is widely distributed, and modern topography is broken by numerous beach ridges, spits and fan-like bars (Ratas, 1976; Eltermann, 1993b). Lake shores are peaty, and the catchment is paludified and mostly forested by pine. Two ditches drain into the lake and one outlet ditch to Tihu Suurjärv. Earlier studies on the Tihu lakes include those of Thomson (1929), Mäemets (1977) and Saarse (1994). The presently studied sediment core was taken from the SE paludified part of the lake (58°51'48"N; 22°32'24"E). The nearest dated section is from Vanajõe (Fig. 1B), 7 km NW of Tihu, where the Litorina Sea mollusc shells from ca. 10–12 m a.s.l. were dated to 3050 yr BP using the electron spin resonance (ESR) method (Molodkov and Raukas, 1996).

Lake Prassi is a small seepage lake (7.5 ha and 7.2 m a.s.l.) in the southern part of Hiiumaa (Fig. 1B), where Silurian limestone is covered by glacial and marine deposits. A 13.5 km long, north-south orientated esker ridge at 8–10 m a.s.l. is situated 500 m east of the lake. The lake itself is surrounded by a small forested mire, and the threshold lies at 6.8 m a.s.l. in the southwestern part. The base of the highest levelled beach ridge is at 10 m a.s.l. (Ratas, 1976). The core was taken from the overgrown eastern part of the lake (58°43'44"N; 22°37'02"E). The nearest locality dated by the ESR method is the Muda Quarry, 4 km north of Prassi (Fig. 1B). It exposes sand rich in mollusc shells, at ca. 8–9 m a.s.l., where the bivalve *Cerastoterna glaucum* was dated to 2700 yr BP (Molodkov and Raukas, 1996).

METHODS

A series of overlapping (0.2–0.5 m) cores were obtained with a Russian peat sampler from the Loopsoo, Tihu and Prassi basins in summer 2012 and from Kõivasoo in summer 1999

(Table 1). One metre long core sections were described in the field, photographed, sealed in plastic liners and transported to the laboratory. Loss-on-ignition (LOI), grain-size, magnetic susceptibility (MS), radiocarbon dates and diatom analyses were carried out on the sediments. Organic matter (OM) was examined continuously from 1 cm thick samples ignited at 525°C for 4 hours, and the results are expressed in percentages of dry matter. The percentage of carbonates (CaCO₃) was calculated after combustion of LOI residue for 2 hours at 900°C. The amount of residue was described as mineral matter and calculated from the sum of organic and carbonate compounds. Volume specific MS κ , expressed in SI units, was measured with a Bartington Instruments MS2E scanning sensor at 1 cm resolution from the carefully cleaned sediment surface. Sediment grain size was measured by a *Horiba* laser scattering particle size analyser from the mineral portion of sediment with the interval of 2.5–5.0 cm. Organic matter was removed by wet oxidation with 30% hydrogen peroxide and carbonates by 10% HCl, and the grain-size classification follows the Udden-Wentworth scale (Last, 2001).

Diatom preparation followed techniques outlined in Battarbee (2001). Diatom samples were digested in hydrogen peroxide and permanently mounted onto microscope slides using Naphrax medium. Usually, about 400 diatom valves were counted in each sample and identified to the species level in order to estimate the percentage abundance of each taxon. In some sandy samples, diatom preservation was poor, but nonetheless, a minimum of 100 identifiable diatoms were counted. Diatoms were grouped according to their living habitat into planktonic, small-sized fragilarioid and periphytic taxa, and with regard to their salinity tolerance into marine/brackish, halophilous, small-sized fragilarioid taxa with brackish-water affinity, small-sized fragilarioid taxa, indifferent, freshwater and unidentified taxa. Diatom floras, used for the identification and ecological information, were derived from different sources

(Krammer and Lange-Bertalot, 1986, 1988, 1991a, b; Snoeijs, 1993; Snoeijs and Vilbaste, 1994; Snoeijs and Potapova, 1995; Snoeijs and Kasperovičienė, 1996; Snoeijs and Balashova, 1998).

The age-depth control of sediment sequences was provided from peat, gytja and plant macrofossils (Table 2). Four dates from charcoal particles, collected by archaeologists from the settlement sites near Kõivasoo, were also considered in the water level curve reconstruction. The material for dating was picked up based on the lithological boundaries, and prioritizing the discovery of terrestrial macrofossils. Macrofossils were extracted by soaking 1 or 5 cm thick samples in water and Na₄P₂O₇ solution, and by wet sieving the material through a 0.20 mm mesh. Obtained terrestrial material was dried at 70°C and dated in the Poznań Radiocarbon Laboratory. The radiocarbon ages were calibrated to calendar years (cal yr BP, 0 = 1950) at 95.4% probability range using the IntCal13 calibration dataset (Reimer et al., 2013) and the *OxCal 4.2* program (Bronk Ramsey, 2009). Radiocarbon dates were combined with lithological data using the OxCal deposition model (Bronk Ramsey, 2008), and weighted average ages were used in the current study (Table 2). Two radiocarbon dates from Loopsoo: 1240 ± 30 yr BP (Poz-50760) and 5140 ± 40 yr BP (Poz-50763), and two dates from Tihu: 1120 ± 40 yr BP (Poz-50761) and 2960 ± 100 yr BP (Poz-52922) (Table 2) are too young and have not been considered in the age-depth model. Diatom and LOI results were plotted using the *TGView* software (Grimm, 2011).

Lithological (Table 1 and Fig. 2), biostratigraphical (Fig. 3) and geochronological proxies (Table 2) of the current and earlier studies (Kents, 1939; Sepp, 1974; Sarv et al., 1982; Raukas et al., 1992, 1996; Molodkov and Raukas, 1996; Königsson et al., 1998; Saarse et al., 2000, 2009) and GIS-based water level surfaces (Saarse et al., 2003, 2006) were used to reconstruct shore displacement curves for Hiiumaa Island. The GIS-based water level surfaces are presented with ±1 metre error bars for

Table 1

Sediment description of studied sites

Site name and altitude	Position of coring site	Depth [cm]	Sediment description
Kõivasoo, 27.5 m a.s.l.	58°54'32"N, 22°11'56"E	0–197	peat
		197–240	coarse detritus gytja
		240–265	calcareous silt with plant remains
		265–318	calcareous silt with mollusc shells
		318–325	silt with mollusc shells
Loopsoo, 21.5–22 m a.s.l.	58°53'41"N, 22°40'21"E	325–351	fine-grained sand with mollusc shells
		0–290	<i>Sphagnum</i> peat, brown
		290–350	transitional peat, dark brown
		350–371	fen peat, light brown
		371–395	minerogenic gytja
Tihu Keskjärv, 14.5 m a.s.l.	58°51'48"N, 22°32'24"E	395–443	sandy silt with organic matter, grey
		443–468	sand
		0–140	peat, well-decomposed, brown
		140–191	coarse detritus gytja, dark brown
		191–204	sand, greyish brown
Prassi, 7.5 m a.s.l.	58°43'44"N, 22°37'02"E	204–206	sand with gravel and pebbles, dark grey
		206–225	silty clay with clasts (till?), bluish-grey
		0–110	reed peat, brown
		110–117	sandy peat
		117–174	sand, fine-grained, grey
174–176	gravel, brownish-grey		
176–200	clay, bluish-grey		
200–225	clay, beige		

Table 2

Radiocarbon dates calibrated at 95.4% probability from the Hiiumaa Island sequences

Site name	Depth, cm/elevation [m a.s.l.]	¹⁴ C date [BP]	Calibrated age, BP (weighted average)	Laboratory no	Dated material	Reference
Kõivasoo	213–223	6580 ± 60	7430–7580 (7510 ± 50)	TA-527	gyttja	Königsson et al. (1998)
Kõivasoo	245.5	6830 ± 90	7750–8000 (7890 ± 60)	Ua-12071	plant remains	Königsson et al. (1998)
Kõivasoo	245–255	7440 ± 60	8020–8250 (8120 ± 70)	TA-528	carbonate fraction	Königsson et al. (1998)
Kõivasoo	260–270	7850 ± 70	8410–8610 (8490 ± 50)	TA-529	carbonate fraction	Königsson et al. (1998)
Kõivasoo	315–325	8190 ± 90	9170–9490 (9350 ± 90)	TA-530	carbonate fraction	Königsson et al. (1998)
Loopsoo	365–366	5190 ± 40	5430–5570 (5490 ± 40)	Poz-52916	peat	current study
Loopsoo	370–371	5050 ± 40	5450–5570 (5520 ± 30)	Poz-52917	peat	current study
Loopsoo	377–382	5600 ± 40	5890–6100 (5980 ± 60)	Poz-52918	plant remains	current study
Loopsoo*	419	1240 ± 30	1070–1270 (1180 ± 60)	Poz-50760	wood (root?)	current study
Loopsoo*	450–455	5140 ± 40	5750–5990 (5670 ± 70)	Poz-50763	plant remains	current study
Tihu	186	4470 ± 40	4900–5270 (5080 ± 90)	Poz-52920	twig	current study
Tihu*	190–195	1120 ± 40	940–1170 (1030 ± 60)	Poz-50761	bark	current study
Tihu	200–204	4490 ± 40	5050–5300 (5200 ± 70)	Poz-52921	plant remains	current study
Tihu*	210–215	2960 ± 100	2870–3370 (3120 ± 130)	Poz-52922	plant remains	current study
Prassi	110	1020 ± 30	830–1050 (940 ± 30)	Poz-52915	plant remains	current study
Prassi	114	1550 ± 30	1310–1430 (1370 ± 30)	Poz-50758	wood	current study
Prassi	123	1480 ± 40	1350–1510 (1410 ± 50)	Poz-50759	bark	current study
Prassi	139	1490 ± 30	1360–1520 (1420 ± 50)	Poz-52914	wood	current study
Prassi	189	1630 ± 30	1420–1600 (1520 ± 50)	Poz-52913	wood	current study
Pihla Bog	505–515/ 14 m a.s.l.	3280 ± 180	3040–3990 (3530 ± 230)	TA-29	woody peat	Liiva et al. (1966)
Kõpu I	26.5 m a.s.l.	5700 ± 70	6320–6660 (6500 ± 80)	TIn-1901	charcoal	Kriiska and Lõugas (1999)
Kõpu I	26.5 m a.s.l.	5330 ± 90	5930–6290 (6110 ± 110)	TA-1493	charcoal	Kriiska and Lõugas (1999)
Kõpu IV	28 m a.s.l.	6760 ± 50	7520–7680 (7620 ± 40)	TIn-2016	charcoal	Kriiska and Lõugas (1999)
Kõpu IV	28 m a.s.l.	6640 ± 60	7430–7610 (7520 ± 50)	TA-2533	charcoal	Kriiska and Lõugas (1999)
Kõpu VIII	28 m a.s.l.	6170 ± 50	6940–7240 (7070 ± 70)	TIn-2024	hazelnut shells	Kriiska and Lõugas (1999)

Dates marked by asterisk have not been considered in the age-depth model

Kõivasoo and Prassi, but not for Loopsoo and Tihu, to keep Figure 4 readable. Timing and changes in the water level at Kõivasoo is based on Saarse et al. (2000), and that of Prassi follows the bio- and chronostratigraphical evidence from the Vääna lagoon in northern Estonia (Saarse et al., 2009), locating at the same Litorina Sea isobase as Prassi.

The GIS-based palaeogeographic maps were created by removing interpolated water level surfaces and thickness of peat deposits from the digital terrain model (DTM; Rosentau et al., 2009). A 10 × 10 m grid-size DTM was used for the current study. The land area is based on the Light Detection And Ranging data (LIDAR) from the Estonian Land Survey. For the off-

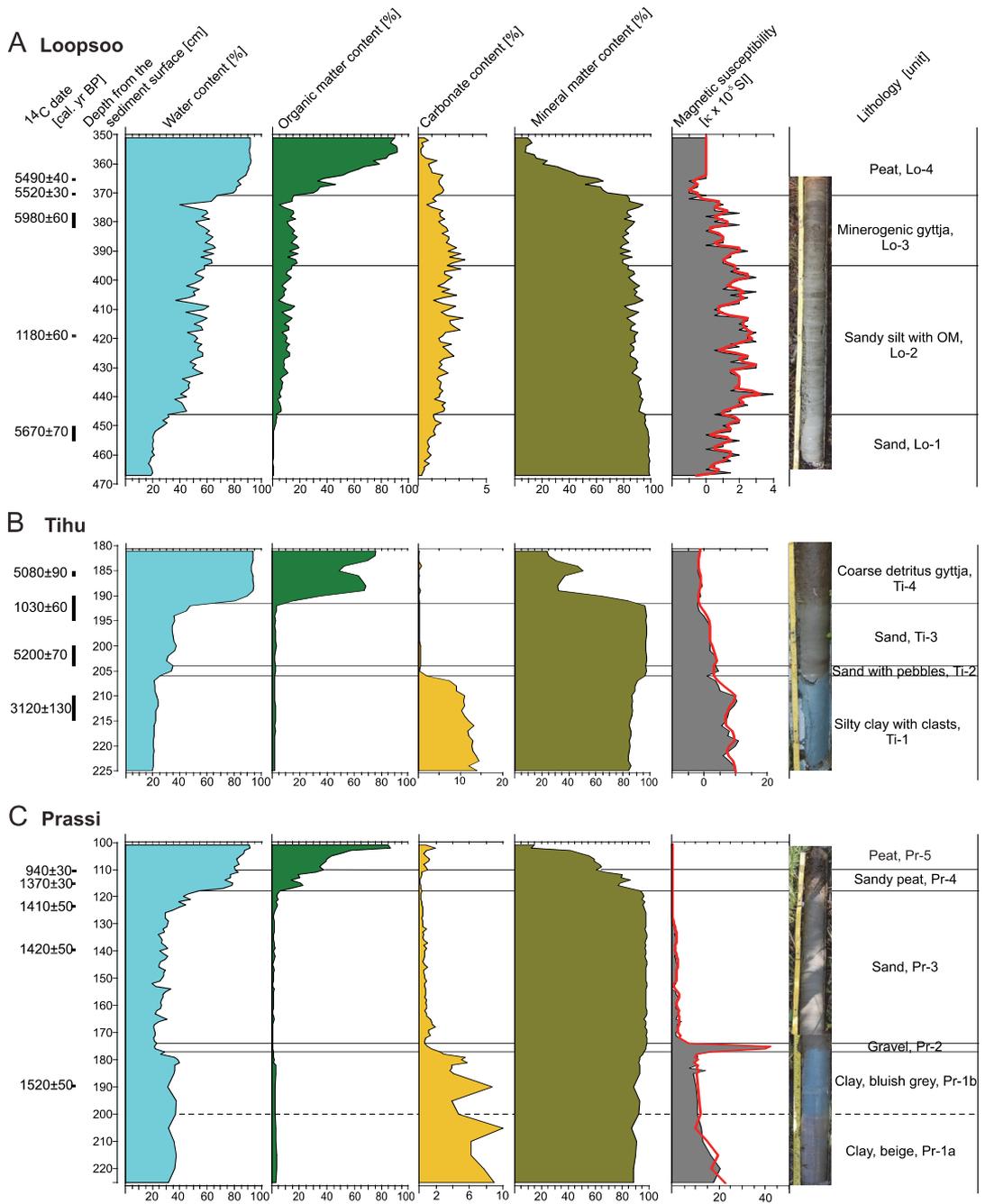
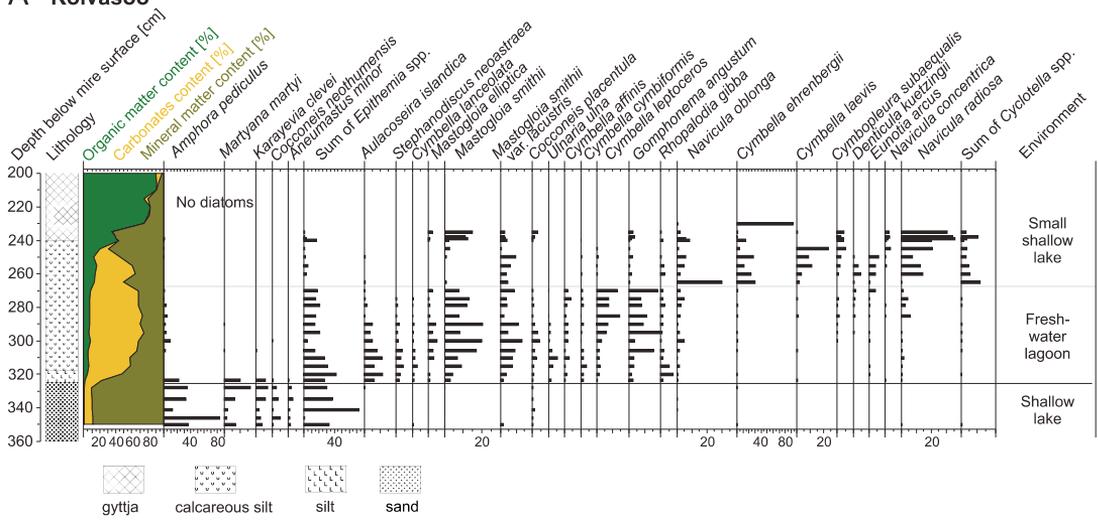


Fig. 2. Loss-on-ignition, magnetic susceptibility results and the lithological units of Loopsoo (A), Tihu (B) and Prassi (C)

MS of Prassi between core depths of 174–176 cm is reduced ten times

A Kõivasoo



B Loopsoo

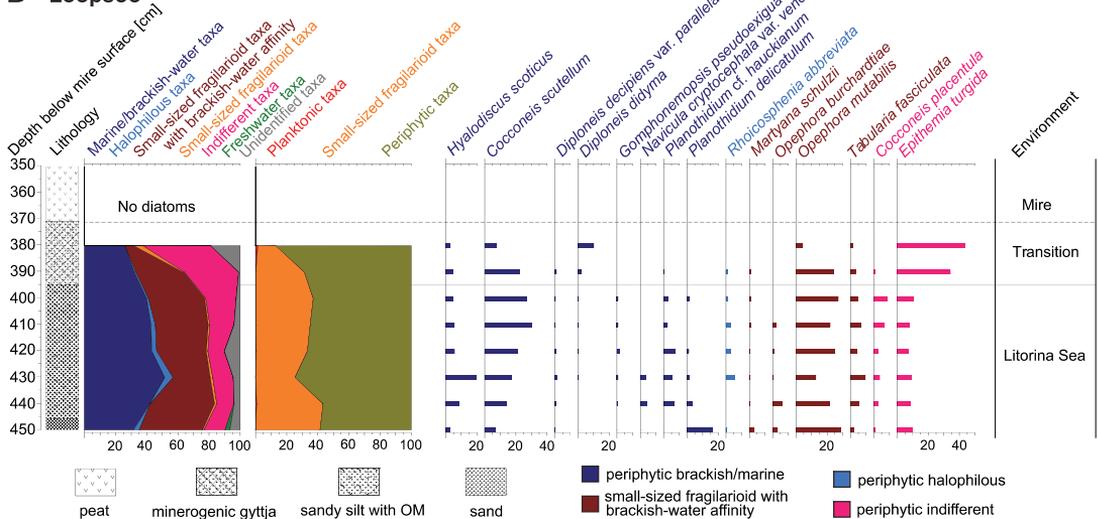


Fig. 3. Diatom diagrams of the Kõivasoo (A), Loopsoo (B),

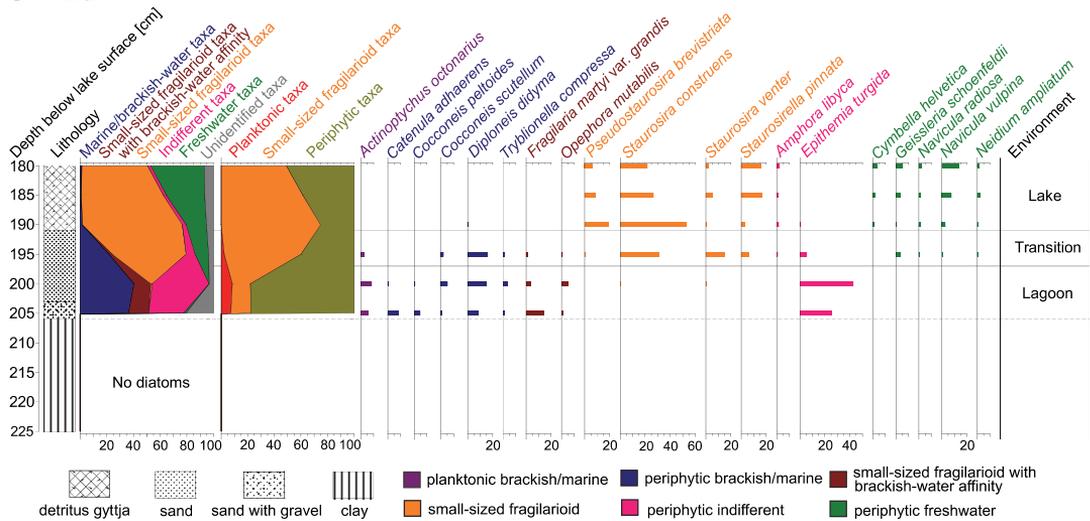
shore area, 1:10,000, 1:25,000 and 1:50,000 scale topographic maps were used. The peat deposits were removed from the DTM according to the 1:10,000 scale soil maps (Estonian Land Survey) and unpublished reports from the Estonian Geological Survey. GIS-based water level surfaces for the Ancylus Lake (10,300 cal. yr BP) and the Litorina Sea (7800 cal. yr BP) were derived from the Estonian coastal formation database (Saarse et al., 2003, 2006). The water level surfaces were created with ±1 metre residual, so that the reconstructed mean water level can fluctuate about 2 metres. In the current paper, the Litorina Sea water level surface (7800 cal. yr BP) was modified according to the assumption that the relative sea level has regressed

evenly due to a linear land uplift (Mörner, 1979; Lindén et al., 2006) and new water level surfaces for 8500, 7100, 6800, 5100, 4800, 4400, 2700 and 2200 cal. yr BP were interpolated.

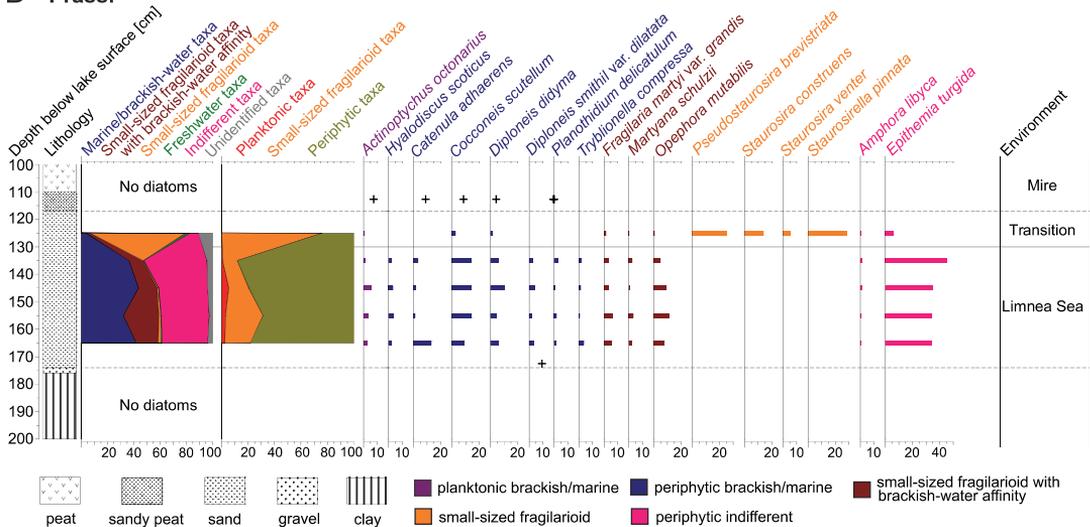
RESULTS

The Kõivasoo sequence (Table 1, Figs. 1B and 3A) contains fine-grained sand with mollusc shell fragments (325–351 cm), silt with mollusc shells (318–325 cm), calcareous silt with mollusc shells (265–318 cm), calcareous silt with

C Tihu



D Prassi



Tihu (C) and Prassi (D) sites

plant remains and mollusc shells (240–265 cm), coarse detritus gytjtja (197–240 cm) and peat from 197 cm upwards (Sarv et al., 1982; Saarse et al., 2000; Table 1). Unlike the other studied sites, silt is highly calcareous and contains over 50% of carbonates. The boundary between calcareous silt and coarse detritus gytjtja is richly paved with mollusc shells.

The diatom flora in the basal part of the Kõivasoo sequence (Fig. 3A) consists of freshwater taxa typical of shallow coastal lake, including *Martynia pedicularis*, *Martynia martyi*, *Karayevia clevei*, *Cocconeis neothumensis* and *Epithemia* spp., which inhabit mostly littoral areas of hard-water lakes. Planktonic and large-lake taxa are absent. A change from shallow-lake taxa to planktonic large-lake taxa, such as *Aulacoseira islandica*, *Stephanodiscus neoastraea* and littoral *Mastogloia* spp. (*Masto-*

gloia elliptica, *M. smithii* and *M. smithii* var. *lacustris*) at a depth of 325 cm refers to the Early Litorina Sea sub-stage and the formation of a freshwater lagoon. Between core depths of 230 and 265 cm, large-lake diatoms disappear and the proportion of *Mastogloia* taxa decreases, being replaced by *Cymbella ehrenbergii*, *C. laevis*, *Navicula radiosa* and other shallow-water small-lake taxa.

The Loopsoo section includes four units: sand, laminated sandy silt with dispersed OM, minerogenic gytjtja, and peat (Table 1 and Fig. 2A). The fine-grained sand layer (Lo-1; 446–468 cm) includes 89–96% of sand, 4–11% of silt and very few OM (Fig. 2A). It is overlain by distinctly bedded sandy silt with dispersed OM (Lo-2; 395–446 cm), containing 53–72% of silt and 28–47% of sand. The content of OM, carbonates and

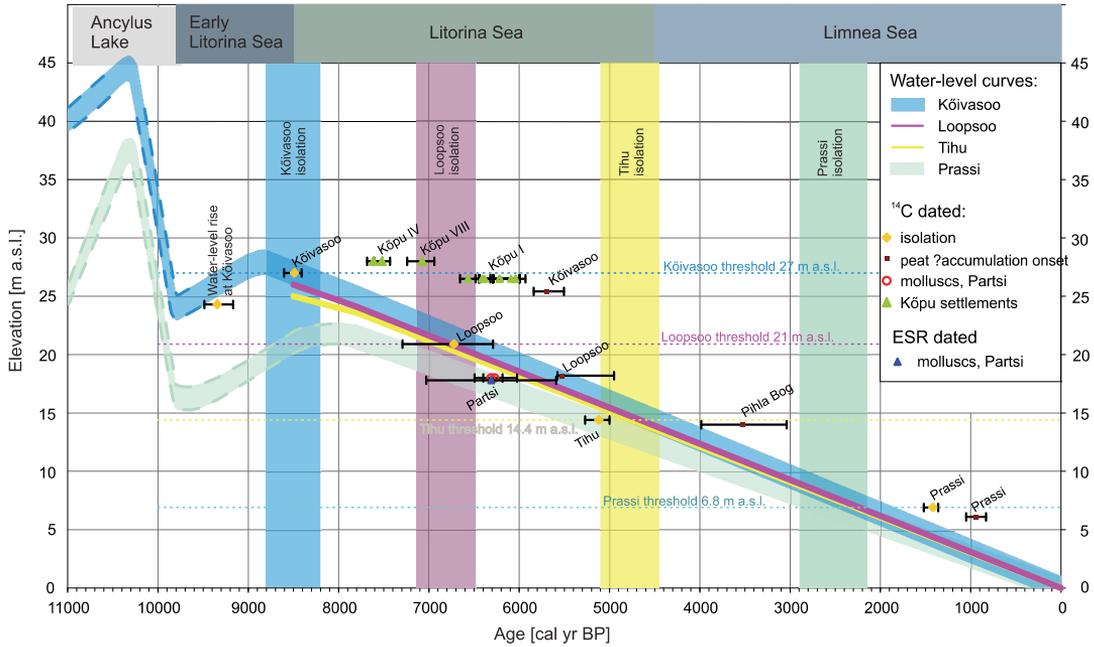


Fig. 4. Water-level curves for the Hiiumaa region, based on different proxies

mineral matter is variable, depending on sediment structure, but still fluctuating in a small range. In minerogenic gyttja (unit Lo-3; 371–395 cm), the OM content slightly rises to 18%, however, the mineral component is still as high as 77–94% (Fig. 2A), of which sand fraction covers 38–53% and silt 47–62%. A major shift in LOI results occurs at 371 cm, where the OM content rapidly increases marking the onset of peat deposition at about 5520 ± 40 cal yr BP (Table 2 and Fig. 2A). The carbonate content remains low throughout the sequence (less 5%), as does MS, fluctuating between $0\text{--}4 \times 10^{-5}$ SI (Fig. 2A).

The diatom composition in the Loopsoo sequence indicates two types of environment, the Litorina Sea and pre-isolation transitional phase (Fig. 3B). Marine/brackish-water and small-sized fragilarioid taxa with brackish-water affinity dominate in the basal sequence to a depth of 395 cm. The sandy layer from the bottom part reveals the highest peak of epipsammic marine/brackish-water *Planothidium delicatulum* (17%) and small-sized fragilarioid *Opephora mutabilis* (29%), indicating a brackish-water environment. Epiphytic diatoms, such as marine/brackish-water *Hyalodiscus scoticus*, *Cocconeis scutellum*, *Gomphonemopsis pseudoexigua*, *Tabularia fasciculata* and halophilous *Rhoicosphenia abbreviata*, which are typically found in the Baltic Sea (Witkowski, 1994; Snoeijns and Balashova, 1998; Witkowski et al., 2000) together with epipsammic diatoms (*Planothidium* cf. *hauckianum*, *Opephora burchardiae* and *O. mutabilis*), comprise about 75% of all identified diatoms. At a core depth of 395 cm, where sandy silt is replaced by minerogenic gyttja, indifferent epiphytic *Epithemia turgida* prevails over marine/brackish-water and small-sized fragilarioid taxa with brackish-water affinity, marking the beginning of the transition zone.

The Tihu sequence consists of four lithological units (Table 1 and Fig. 2B): bluish-grey silty clay with clasts resembling waterlain till (Ti-1, 206–225 cm), sand with gravel and pebbles typical of erosional surfaces (Ti-2, 204–206 cm), sand (Ti-3,

191–204 cm), and coarse detritus gyttja (Ti-4, 140–191 cm). The OM content in basal silty clay is 1–2%, the carbonate content fluctuates between 9–14%, mineral matter is up to 91%, and MS up to 10×10^{-5} SI (Fig. 2B). Mineral matter is dominated by the clay fraction, accounting for 40–51%. The silt and sand contents vary between 39–46% and 10–14%, respectively. In addition, this sediment section contains sparse gravel grains and pebbles. Macrofossils from a core depth of 210–215 cm are dated to 3120 ± 130 cal yr BP (Table 2), which is not consistent with both the uppermost date and the suggestion that this sediment is waterlain till.

The erosional bed at 204–206 cm (Ti-2) consists of sand with gravel and pebbles, cemented sandstone nodules and broken mollusc shells. Its LOI results are similar to those of the overlying sand (Fig. 2B). Sand (Ti-3; 191–204 cm) is fine to very fine in grain size with the OM content less than 3%, carbonates 0.2–0.6%, and mineral matter up to 97%. The MS values are low and range between $1\text{--}4 \times 10^{-5}$ SI (Fig. 2B). According to grain size distribution, the sand fraction content fluctuates between 78–94%, and silt between 6–29%, whereas clay fraction is absent. Deposition of this sediment started about 5200 ± 70 cal yr BP. At 191 cm (Ti-4), sand is replaced by coarse detritus gyttja. The content of OM increases to 76%, and the content of mineral matter decreases to 24% (Fig. 2B). The AMS date from bark at the isolation contact (190–195 cm) shows an age of 1030 ± 60 cal yr BP (Table 2) and probably represents a macrofossil remain that is swept downwards during coring. If the radiocarbon dates from coarse detritus gyttja and the basal part of sand are correct (Table 2), then the 12 cm thick sand (unit Ti-3) was deposited during a period of 60–200 years.

No diatoms have been found in unit Ti-1 (Fig. 3C), supporting the interpretation that the unit represents waterlain till. The first remnants of broken diatom frustules were observed in the sandy and gravelly unit Ti-2 at a depth of 205 cm. As the preservation of diatoms in this layer was poor, broken parts of marine/brack-

ish-water planktonic *Actinopterychus octonarius*, epipellic *Diploneis didyma* and indifferent epiphytic *Epithemia turgida*, and small-sized diatoms resistant to erosion, such as *Fragilaria martyi* var. *grandis* and *Catenula adhaerens*, have been identified. The dominance of marine/brackish-water planktonic *Actinopterychus octonarius* (8%), epiphytic *Cocconeis scutellum* (5%), epipellic *Diploneis didyma* (15%) and *Tryblionella compressa*, and indifferent epiphytic *Epithemia turgida* (43%) at a depth of 197–205 cm, i.e. representing unit Ti-2 and the lower part of unit Ti-3, indicate a brackish-water environment and most likely a pre-isolation lagoonal phase of the basin. The short-lived transition stage is characterized by mass occurrence of small-sized fragilarioid taxa such as *Staurosira construens* (up to 50%), *S. venter* (15%) and *Staurosirella pinnata*, decreased abundance of marine/brackish-water and indifferent taxa, and appearance of freshwater periphytic diatoms like *Geissleria schoenfeldii*, *Navicula vulpina*, *Neidium ampliatum* etc.

The lithology of Prassi (Table 1 and Fig. 2C) is similar to that of Tihu. The basal clay unit is subdivided into beige (Pr-1a, 200–225 cm) and bluish-grey (Pr-1b, 176–200 cm) clayey sub-units covered by a thin gravel bed (Pr-2, 174–176 cm), sand (Pr-3; 117–174 cm) and sandy peat (Pr-4, 110–117 cm) that gradually turns upwards to peat (Table 1 and Fig. 2C). The OM content of basal clay remains between 1.4–3.8%, but carbonates reach 10%. The grain size of differently coloured clay is quite similar and composed mostly of clay (67–89%) and silt (11–27%) fraction, with subordinate sand fraction. A woody piece found at a core depth of 189 cm was AMS-dated to 1520 ± 50 cal yr BP (Table 2). The overlying gravel (Pr-2) is poor in OM and carbonates but rich in magnetic minerals, up to 400×10^{-5} SI (Fig. 2C). Sand (Pr-3) contains a very low amount of OM, carbonates and magnetic minerals (Fig. 2C). Its grain size distribution is quite stable, with 75–87% of sand fraction and 13–25% of silt fraction. In sandy peat (Pr-4), at a depth of 100 cm, the OM content increases to 85%.

The basal clay, gravel and sand do not contain any diatom valves to a depth of 165 cm (Fig. 3D). The upper part of the sequence (110–117 cm) is also barren of diatoms. Diatom frustules have been identified and counted in the interval 120–165 cm. Like in the Loopsoo sequence, the diatom assemblage of the Prassi sequence is dominated by epiphytic marine/brackish *Hyalodiscus scoticus* and *Cocconeis scutellum* (14%), indifferent *Epithemia turgida* (35–46%) and marine/brackish-water epipsammic *Catenula adhaerens* (13%), *Fragilaria martyi* var. *grandis* (6%), *Martyana schulzii*, *Opephora mutabilis* (11%), all characteristic of the brackish shallow coastal waters (Witkowski, 1994). Abundance of marine/brackish-water planktonic *Actinopterychus octonarius* indicates that the Prassi Basin is a shallow, open bay of the Limnea Sea. The sharp change of diatom assemblages from marine/brackish-water and indifferent taxa to small-sized fragilarioid taxa (*Pseudostaurosira brevistriata*, *Staurosira construens*, *S. venter* and *Staurosirella pinnata*) that prefer freshwater conditions, marks the onset of the transition phase at 130–135 cm and the onset of isolation from the BSB.

DISCUSSION

LITHO-, BIO- AND CHRONOSTRATIGRAPHY

Basal sand (325–351 cm) from Kõivasoo contains mostly periphytic epipsammic (taxa attached to sand and silt grains) freshwater diatoms (e.g., *Amphora pediculus*) and indifferent

Epithemia spp. (Fig. 3A) inhabiting the bottom of shallow hard-water lakes in Estonia. Such a diatom assemblage refers to a low water level and an isolated shallow lake environment before 9400 cal yr BP. The water level dropped below the Kõivasoo threshold. Its exact level at that time is unknown, as it was filled with sand during the later transgression. Changes in the lithology at 325 cm (Table 1), a sharp decrease in *Pediastrum* (Königsson et al., 1998), and the appearance of large-lake planktonic diatoms (Fig. 3A) and molluscs, preferring deeper water environments (Kessel and Raukas, 1967), suggest that the Kõivasoo Basin was connected with the BSB 9350 ± 90 cal yr BP (Table 2), and the water level was higher than 24.3 m a.s.l. The diatom composition indicates that a freshwater lagoonal environment (Fig. 3A) with calcareous deposition (325–265 cm) lasted until ca 8490 ± 50 cal yr BP in Kõivasoo (Table 2). Disappearance of large-lake diatoms, decrease in *Mastogloia* spp. and replacement by taxa characteristic of small shallow lakes between core depths of 265 and 230 cm represent the final isolation of Kõivasoo from BSB.

The diatom flora shows that the sandy-silty beds in the Loopsoo Basin were deposited in the Litorina Sea. The presence of epipsammic (*Planolithidium delicatulum*, *P. cf. hauckianum*, *Martyana schulzii*, *Opephora mutabilis*) and epiphytic (*Hyalodiscus scoticus*, *Cocconeis scutellum*, *C. placentula*) brackish-water diatoms indicates a shallow depositional environment (Fig. 3B). According to a shoreline displacement simulation, the water level near Loopsoo was at ca. 22 m a.s.l. by the onset of deposition of minerogenic gytjtja (unit Lo-3), holding only ca. 1.5–2.0 m above the threshold elevation, and promoting erosion of the surrounding beach ridges and influx of sand to the Loopsoo Basin during the higher wave energy. The onset of isolation was most probably gradual and is recorded between 395 and 380 cm in the sediment section, which suggests that organic production increased prior to the final isolation of the basin (Lindén et al., 2006). According to the age-depth model, isolation started about 6720 ± 270 cal yr BP and terminated 5980 ± 80 cal yr BP, covering a transition period when marine/brackish-water diatoms still inhabited the basin (Fig. 3B). However, the long-lasting isolation is in conflict with water level simulation results, which show that isolation terminated about 6500 cal yr BP. Such discrepancy could be explained by proximity of the Loopsoo Basin to the sea and location on a small island exposed to the winds and wave activity. Considering that the sedimentation of 24 cm thick minerogenic gytjtja lasted 1200 years and the sedimentation rate was 0.2 mm yr^{-1} , a gap between gytjtja and peat seems to be also realistic. In the upper part of minerogenic gytjtja, where diatoms are absent, the sediment grain size turns more sandy due to shallowness of the basin and increased erosion.

The basal bluish-grey silty clay with clasts (unit Ti-1) in the Tihu sequence is most probably waterlain till widespread on Hiiumaa Island (Eltermann, 1993a, b; Kadastik and Kalm, 1998; Kalm and Kadastik, 2001; Kadastik, 2004). These deposits are also found in the bottom of Tihu Suurjärv (Saarse, unpubl.). The grain size distribution chart of waterlain till is multimodal and the contact with the overlying beds is sharp and paved with dropstone (Kadastik and Kalm, 1998). If the basal silty clay is waterlain till, there should be a long-lasting hiatus between sand and clay, as it deposited not later than 13,000 years ago (Saarse et al., 2012), and mid-Holocene sand during the Litorina Sea stage. However, it is difficult to explain why the older portion of the Litorina Sea sediment is absent. Location of studied core on the basin slope, accessible to erosion of stormy waves, could be a rationale. Sand with gravel and pebbles between clayey and sandy deposits indicate a water level lowering

and erosion. A question arises, when such erosion took place. If that basal clay deposited during the Ancylus Lake, and considering that the Ancylus Lake maximum level at Tihu was 41 m a.s.l. and that the water level drop during the following regression was 30 m (Raukas and Ratas, 1996), the Tihu area should have been emerged, which would explain the erosion and hiatus. Thus, the basal silty clay with clasts represents most likely waterlain till.

During the Litorina Sea stage, typical marine diatoms (*Catenula adhaerens*, *Diploneis didyma*, *Actinopterychus octonarius* etc.) inhabited the basin (Fig. 3C). Dominance of epiphytic and epipsammic diatoms indicates that, Tihu, like Loopsoo, was a shallow overgrowing lagoon of the Litorina Sea at that time. However, unlike at Loopsoo and Prassi, the transition zone from a lagoon to a freshwater basin in Tihu is marked by a peak of small-sized fragilarioid taxa (Fig. 3C), indicating a change in depositional environment (Seppä and Tikkanen, 1998; Seppä et al., 2000; Risberg et al., 2005; Grudzinska et al., 2012, 2013). Sand replacement by coarse detritus gyttja marks the isolation contact at 191 cm and is in accordance with a change in diatom assemblage from marine/brackish-water taxa to freshwater taxa (Fig. 3C).

Similarly to the Tihu section, the sediment sequence of Prassi also includes lowermost clayey deposits covered by a thin gravel bed rich in magnetic minerals (Fig. 2C). The basal clay can be of glaciolacustrine origin (Eltermann, 1993a), as suggested by the absence of diatoms. However, the AMS radiocarbon date indicates a much younger age (Table 2) and refers to re-deposition of material and/or contamination during coring. The overlying gravel bed with a high content of magnetic minerals is interpreted as an erosional event. If the basal clay is re-deposited and the AMS date 1520 ± 50 cal yr BP is reliable, then the erosional surface was formed about 1440 ± 50 cal yr BP. Marine/brackish-water diatoms, characteristic of the Limnea Sea, are preserved only in the middle part of sand (Fig. 3D), whereas the lower and upper parts of sand were barren of diatoms. According to the diatom composition, the beginning of the transition from the Limnea Sea to a more or less isolated water body is recorded at a core depth of 135 cm (Fig. 3D), but the absence of diatoms in the upper part of the sequence makes determination of the exact isolation contact indistinct. According to ^{14}C dates, Prassi isolated about 1400 cal yr BP, but this age is in conflict with the water level simulation results of the present study, which show that the isolation occurred about 2700 cal yr BP. This age is concurrent with the ESR date from *Cerastoterma glaucum* of the Muda Quarry, 4 km north of Prassi (Molodkov and Raukas, 1996).

RELATIVE SEA LEVEL CHANGES AND DEVELOPMENT OF HIIMUMAA

Relative sea level curves compiled for Kõivasoo, Loopsoo, Tihu and Prassi are presented in Figure 4. According to the GIS-based water level surfaces, derived from the Estonian coastal formation database (Saarse et al., 2003), the mean relative water level during the Ancylus Lake transgression about 10,300 cal yr BP was 44.5 m a.s.l. at Kõivasoo, 41.6 m at Loopsoo, 40.9 m at Tihu, and 37.5 m at Prassi (Fig. 4). During the Litorina Sea transgression 7800 cal yr BP, the mean water level was 24.9 m a.s.l. at Kõivasoo, 24.1 m at Loopsoo, 23.6 m at Tihu, and 21.5 m at Prassi. The reconstructed water level curve shows a linear decline since the Litorina Sea transgression, but it does not rule out minor changes being within error limits of simulation (Fig. 4). If large-scale water level fluctuations would have taken place between ca. 8000 cal yr BP (Kõivasoo)

and ca. 1500 cal yr BP (Prassi), they should have been visible in diatom stratigraphy and lithostratigraphy of the presently studied sections. The reconstructed water level curves display a diachronous Early Litorina/Litorina Sea transgression peak occurring earlier in the Kõivasoo area and later in the Prassi area as a result of the different rate of land uplift. The gradient of the highest Litorina Sea shoreline is ca 16.5 cm km^{-1} between Kõivasoo (at 27 m isobase) and Prassi (at 22 m isobase), and the distance between them is 30 km. The calculated summary land uplift during the last 10,300 years was 4.4 mm yr^{-1} at Kõivasoo, and has decreased to 2.5 mm yr^{-1} at present (Torim, 2004). These results complement observations of recent investigations by Veski et al. (2005), Saarse et al. (2010), Grudzinska et al. (2013) and Rosentau et al. (2013).

The first island to emerge from the BSB about 11,000 cal yr BP was Kõpu, and the first shallow lake to isolate was Kõivasoo (Fig. 5). Kõpu Island, locating 80 km from mainland Estonia and containing abundant sand and gravel deposits, was long time subjected to strong winds and stormy waves that favoured the development of very mosaic topography as well as the Yoldia Sea/Ancylus Lake shoreline at 55–30 m a.s.l. (Kents, 1939) and the Early Litorina shoreline at 28–27 m a.s.l. The next island to emerge about 9000 cal yr BP was a triangular outwash plane called Hiiu Island (Ratas, 1976). By 7800 cal yr BP, these two islands and three tiny islets were the only ones which formed the core of the present-day Hiiumaa (Fig. 6A). Despite the small size and distant location of Kõpu Island from the mainland, it was seasonally colonized by seal hunters between 7600 and 6100 cal yr BP, as confirmed by artefacts and osteological material (Kriiska and Lõugas, 1999). During that time, Kõivasoo already existed as an isolated lake and provided freshwater for hunters enhancing their stay on the island. Shallowing and paludification of Kõivasoo Lake could be one of the reasons why their nearby settlements were abandoned. About 6100 cal yr BP and about 5500 cal yr BP, Kõivasoo was fully overgrown by peat (Sarv et al., 1982).

Before the isolation, Loopsoo underwent a lagoonal (transitional) phase being strongly affected by the sea, as it was located on the small island in an open sea (Fig. 7A). Between

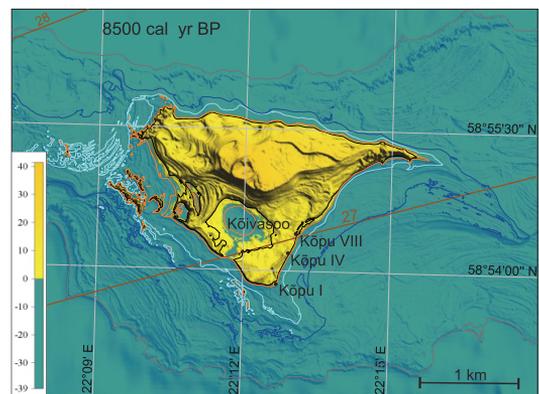


Fig. 5. 3D reconstruction at 8500 cal yr BP at Kõivasoo

Modelled water level surface isobases are indicated by brown lines together with elevations in metres a.s.l.; the shoreline is shown with a modelling error $\pm 1\text{m}$ (black line $+1 \text{ m}$, orange line -1 m); light blue line corresponds to the 5 m and blue to the 10 m water depth; black dots mark the location of archaeological sites

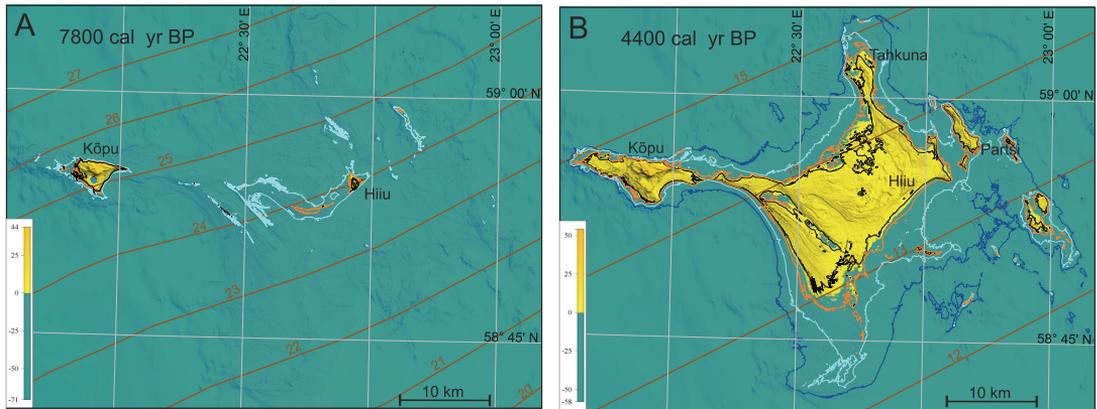


Fig. 6. 3D reconstructions of Hiiumaa Island at 7800 cal yr BP (A) and 4400 cal yr BP (B)

For explanations see Figure 5

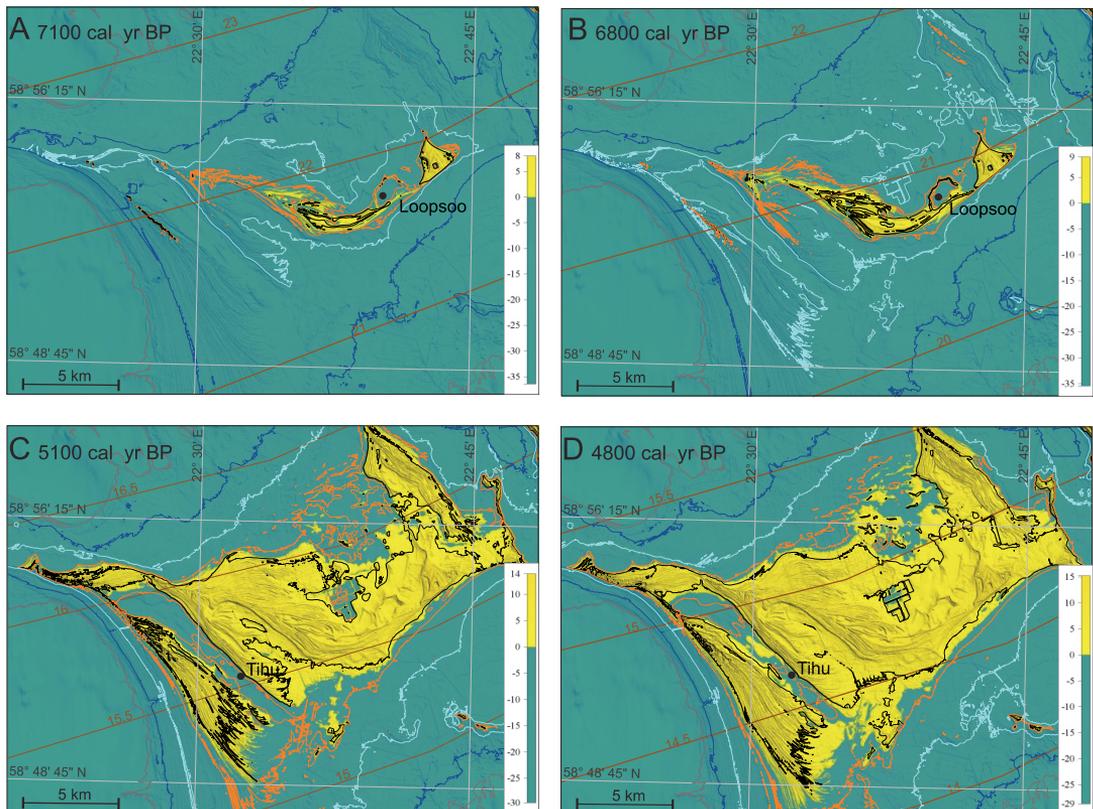


Fig. 7. 3D reconstructions at 7100 and 6800 cal yr BP in the surroundings of Loopsoo (A, B) and at 5100 and 4800 cal yr BP in the vicinity of Tihu (C, D)

For explanations see Figure 5

7100 and 7000 cal yr BP, sandy and shingle ridges surrounded Loopsoo Basin with narrow passage in north-east (Fig. 7A). It began to isolate from the Litorina Sea most likely before 6800 cal yr BP, nevertheless, with storms and high wave activity, sea water flooded the basin during and after the isolation (Fig. 7B).

About 5100 cal yr BP, an elongated shallow lagoon existed in the Tihu depression with a passage to the sea in the south-east (Fig. 7C), which closed approximately 4800 cal yr BP. However, it seems likely that high waters were able to enter into the basin (Fig. 7D). Afterwards, the Tihu depression gradually transformed to land area where Tihu lakes remained as residual ones.

The youngest of the studied basin to isolate was Prassi. A shallow lagoon with a passage in the south occurred in the Prassi area about 2700 cal yr BP (Fig. 8A). A prolonged isolation process of the Prassi site was favoured by a flat topography, slightly inclined towards the west and south, a lack of barrier ridges, and exposure to the open sea (Fig. 8). Considering that the threshold elevation is at 6.8 m a.s.l., Prassi started to isolate about 2500 cal yr BP, which is in conflict with radiocarbon dates (Table 2) that we discussed earlier. Furthermore, it is also possible that the uppermost part of sand with abundant woody pieces have been carried into the basin during heavy storms when the water level may rise up to 3 m, as it was during the January storm in AD 2005.

It can conclude from the 3D reconstructions that Hiiumaa Island is relatively young in the present geographical shape. Its highest peak 68 m a.s.l. emerged from the BSB during the Yoldia Sea stage approximately 11,000 cal yr BP. Emerging BSB processes left behind erosional-prone sandy fields for waves and wind to create tiny islets, beach ridge systems, fan like bars, sandy terraces and spits forming now a very mosaic landscape. By 7800 cal yr BP, only two large islands, Kõpu with an area of ca. 4 km² and Hiiu ca. 1 km², had been emerged (Fig. 6A). Due to a relatively fast land uplift by 4400 cal yr BP, almost half of the present-day island was emerged from the sea (Fig. 6B).

CONCLUSIONS

1. Kõpu Peninsula, the highest part of Hiiumaa Island, started to emerge from the sea during the Yoldia Sea about 11,000 cal yr BP and remained as an isolated island up to the onset of the Limnea Sea. The summary uplift rate since the emergence of Kõpu is 5.6 mm yr⁻¹ which has decreased to 2.5 mm yr⁻¹ at present.

2. According to the GIS-based water level simulation, the mean relative water level during the Ancylus Lake (about 10,300 cal yr BP) was 44.5 m a.s.l. at Kõivasoo, 41.6 m at Loopsoo, 40.9 m at Tihu, and 37.5 m at Prassi, and during the Litorina Sea (about 7800 cal yr BP) accordingly 24.9 m a.s.l. at Kõivasoo, 24.1 m at Loopsoo, 23.6 m at Tihu and 21.5 m at Prassi.

3. Diatom evidence indicates that, before 9400 cal yr BP, the Kõivasoo Basin had existed as a small shallow lake. Afterwards, it was connected with the BSB, and a semi-closed fresh-

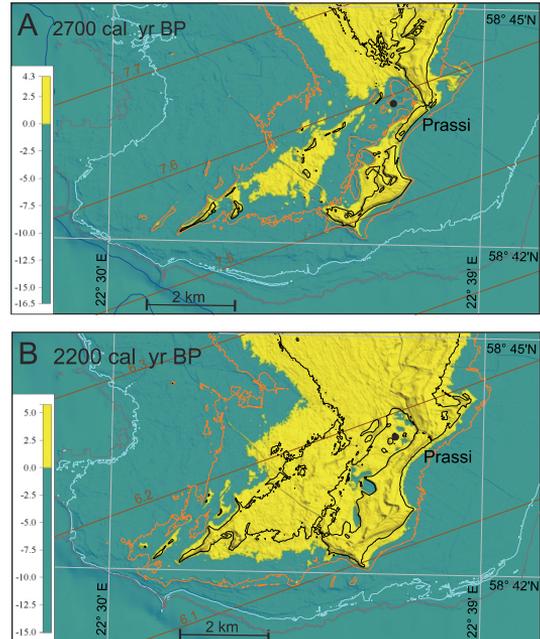


Fig. 8. 3D reconstructions at 2700 and 2200 cal yr BP at Prassi (A, B)

For explanations see Figure 5

water lagoon was formed until about 8500 cal yr BP, when the basin was finally isolated.

4. According to diatom records, the Loopsoo, Tihu and Prassi basins isolated gradually passing the transitional phase at 6800, 4800 and 2500 cal yr BP, respectively.

5. Palaeogeographic reconstructions show that the isolation of the basins lasted about 500–800 years, being longer for the basins that isolated later.

6. GIS-based modelled isolation times are supported by radiocarbon dates, except for Prassi.

7. Water level curves for different sites in Hiiumaa follow a similar pattern, but the transgression sites in Hiiumaa follow a similar pattern, but the transgression maximum in areas of slower land uplift (Prassi) occurred later than in areas of faster uplift (Kõivasoo), showing a diachronous Early Litorina/Litorina transgression peak.

8. Water level curves display a linear decline during the last 8000 years.

Acknowledgements. The study was supported by the institutional research funding IUT 1-8, ESF Grant 9031 and Doctoral Studies and Internationalisation Programme DoRa. We thank two anonymous reviewers for valuable comments and suggestions.

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V Grudzinska, I., Vassiljev, J., Saarse, L., Veski S. (submitted). Diatom and sedimentary record of past environmental change and paleo sea water intrusions into coastal Lake Lilaste, Latvia. *Journal of Paleolimnology*.

Diatom and sedimentary records of past environmental change and paleo sea water intrusions into coastal Lake Lilaste, Latvia

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Keywords Litorina Sea • Diatoms • Loss-on-ignition • Relative sea level • AMS dates • Holocene

Abstract

Diatom and sedimentary records of a 10-m long sediment sequence from coastal Lake Lilaste were analysed to evaluate environmental changes during the Holocene related to past sea water intrusions. Lilaste is located ca. 1 km from the present sea coast in an area with low uplift rates and has a threshold of 0.5 m a.s.l.; thus, it was considered to be an appropriate site to study the influence of sea level fluctuations on sedimentary conditions. According to the age-depth model, sedimentation in the lake started at ca. 11,400±190 cal yr BP, shortly after the Baltic Ice Lake (BIL) drainage. Variations in diatom composition, along with sediment lithostratigraphical interpretation, show that a shallow, nutrient-rich freshwater lake existed during the Early Holocene. First, brackish water diatoms appear concurrently with the Litorina Sea transgression ca. 8700±50 cal yr BP, but long-term intermittent influxes of brackish water were observed between 6700±40 and 4200±80 cal yr BP. In this time span, diatom taxa indicate an increased nutrient content and a high conductivity state explained by occasional mixing of brackish and fresh water that promotes biological productivity. Lilaste was finally isolated from the sea at ca. 4200±80 cal yr BP, after which a stable freshwater environment, dominated by planktonic diatoms such as *Aulacoseira ambigua*, *A. granulata*, *A. islandica* and *A. subarctica*, was established. At 400±50 cal yr BP, planktonic diatoms were gradually replaced by *Fragilaria* spp., indicating the beginning of anthropogenic impact. A reconstructed relative water-level curve is coincident with the eustatic sea level from 6800±40 cal yr BP onwards. At the same time, there was a distinct increase in the abundance of brackish water diatoms, as the sea level reached the threshold of Lilaste, which at that time was probably approximately 3 m below the present sea level. According to diatom composition and interpolated radiocarbon dates, the Litorina Sea transgression culminated almost 1000 years later and was a long lasting event (ca. 2200 years) in areas with a small land uplift rate close to zero, e.g., the southern part of the Gulf of Riga, compared with sites at high isolines in the northern part of the Baltic Sea.

Introduction

Diatom assemblages in lakes depend on different factors, such as temperature, water transparency, turbulence, ice cover length, pH, nutrients and salinity (Battarbee et al. 2001). Due to the ability of diatoms to quickly respond to changes in physical, chemical or biological conditions, and their good preservation in sediments, palaeolimnologists widely apply diatom analyses as indicators of environmental change. Diatom analysis has often been used in isolation basin studies to reconstruct past sea-level fluctuations in areas affected by glacio-isostasy, such as Sweden (Yu et al. 2004; Risberg et al. 2005), Finland (Eronen 2001; Miettinen 2004), Norway (Corner et al. 1999; Balascio et al. 2011), Estonia (Saarse et al. 2009a, 2009b; Grudzinska et al. 2013, 2014), Scotland (Shennan et al. 1996, 2000) and Greenland (Long et al. 2011). It is relatively easy to determine isolation contact, where marine and brackish diatoms are replaced by halophilous and subsequently by freshwater diatoms, to date the sediments and detect the threshold in areas with high land uplift rates and crystalline bedrock. In areas where land uplift is nearly zero and sedimentary sequences are composed of sand, silt and/or clay, determination of threshold and lake isolation time is a rather complicated task. Such an area is the coast of the Gulf of Riga (Fig. 1), where apparent land uplift is approximately 1 mm yr^{-1} in the northern part, and the zero isoline crosses the southern part of the area (Ekman 1996). Due to substantial sediment input from three large rivers, longshore sediment transport and low land uplift rate, sediment stratigraphy in the vicinity of Riga differ considerably. The area has been transformed by several transgressions and regressions of the Baltic Sea. Consequently, it was considered that sediment sequences from coastal lakes could reveal marine water intrusions because of changes in the sea water level during the Holocene.

According to previous studies (Grinbergs 1957; Ulsts 1957), the Coastal Lowland in the southern part of the Gulf of Riga was inundated twice by the Litorina Sea transgression, and two ancient shorelines were formed according to radiocarbon dates (Veinbergs 1979) at $8600\text{--}8000 \text{ cal yr BP}$ ($7750\pm 180 \text{ Riga-192}$ and $7110\pm 170 \text{ Mo-224 }^{14}\text{C BP}$) and approximately 4600 cal yr BP ($4060\pm 50 \text{ }^{14}\text{C BP Tln-22}$). Conclusions regarding the sea level changes in Latvia, however, were mainly based on morphological, lithological and pollen data, with a limited number of radiocarbon dates, and there was no common agreement among researchers whether all lakes in the Coastal Lowland were affected by transgression waters.

The aim of this study is to improve the understanding of the nature and extent of Holocene sea level changes in the eastern Baltic Sea region, in the isostatic hinge area with low land uplift rates, and to elucidate the effect of brackish water on freshwater lake diatom composition. Lake Lilaste was selected for palaeoecological investigation because of its location close to the sea coast, low lying threshold and long continuous sediment sequence. Applied multi-proxy analyses provided evidence regarding 1) whether Lake Lilaste was affected by sea-level changes, 2) the time periods when they occurred, and 3) how the lake environment was altered.

Site description

Lake Lilaste (Fig. 1b; $57^{\circ}10'44.4'' \text{ N}$; $24^{\circ}21'05.9'' \text{ E}$) is located on the Coastal Lowland, Rigavas Plain, approximately 20 km northeast of Riga, and 1 km from the Gulf of Riga at an altitude of 0.5 m a.s.l. The area of the lake is 1.836 km^2 , including the area (0.035 km^2) of one island. The maximum water depth is 3.2 m, with an average of 2.0 m. It is a drainage lake, with inflow from the Melnupe River and outflow *via* the Lilaste River to the Gulf of Riga. Moreover, biologically treated wastewater is discharged to the lake, resulting in an increasing anthropogenic stress. Lake Lilaste has a connection with Lake Dunu, once forming a larger waterbody and lagoon of the BIL. The lake bottom is flat and covered by gyttja, with a maximum thickness of approximately 10 m. The lake shores are shallow, covered by reeds (the name 'Lilaste' in the Liv language means 'reeds'), in

some places sloped, and surrounded by ancient dunes. The lake catchment area is 144 km² and is mostly covered by pine forest and bogs. Devonian sandstone and clay (30–40 m b.s.l.) form the bedrock, which is overlain by ca. 45 m thick Quaternary cover consisting of Weichselian till, glaciolacustrine and marine silt, sand and aeolian deposits. A new residential area with an artificial peninsula expands on the southeastern shore, and on the northwestern shore there is an angler centre, a bathing place, and recreational facilities. Along western shore runs highway A1 Riga–Tallinn and a railway (Fig. 1b).

Materials and methods

In spring of 2012 and 2013, sediment samples from Lake Lilaste were obtained through the ice with a Russian peat sampler in the deepest part of the basin. The uppermost 0.4 m of unconsolidated sediment was sampled by a Willner-type sampler, and slices 2 cm thick were cut and put into plastic bags. Core sections (each 1 m long) were described in the field, photographed, sealed in plastic liners, and stored in a laboratory cold-room.

The lithology of the core was studied by loss-on-ignition (LOI) following Heiri et al. (2001), and magnetic susceptibility (MS) following Nowaczyk (2001). Continuous samples 1 and 2 cm thick were used for LOI analysis. The organic matter (OM) content was measured at 525°C for 4 h, and expressed in percentage of dry matter. The percentage of carbonate content was calculated after combustion of LOI residue for 2 hours at 900°C. The mineral fraction (MM) content was calculated based on the sum of organic and carbonate compounds. Volume specific MS κ , expressed in SI units, was measured with a Bartington MS2E high-resolution scanning sensor at 1 cm resolution from the cleaned sediment surface covered with a thin plastic film.

The samples for diatom analysis were prepared by the following techniques described in Battarbee et al. (2001). To remove OM, samples were digested in 30% hydrogen peroxide. Carbonates, metal salts and oxides were removed by adding 10% hydrochloric acid, and repeated decantation was applied to extract fine and coarse mineral particles. To estimate the concentration of diatoms and to calculate the diatom accumulation rate (DAR), a known quantity of synthetic microspheres was added to diatom suspension. A drop of the remaining residue was spread over the cover slip, dried overnight at room temperature and mounted permanently onto microscope slides with Naphrax resin. For most samples, approximately 500 valves were counted under oil immersion using a Zeiss Axio Imager A1 microscope with differential interference contrast illumination at $\times 1000$ magnification and were identified to species level to estimate the percentage abundance of taxa. Diatoms were grouped according to their salinity tolerance into marine/brackish, halophilous, small-sized fragilarioid taxa with brackish water affinity, small-sized fragilarioid taxa preferring freshwater, indifferent, freshwater, and unidentified taxa. Habitat classification included planktonic, small-sized fragilarioid, and periphytic taxa. Diatom floras used for identification and ecological information were based on well-established sources (Krammer and Lange-Bertalot 1986, 1988, 1991a, 1991b; Snoeijs 1993; Snoeijs and Vilbaste 1994; Snoeijs and Potapova 1995; Snoeijs and Kasperovičienė 1996; Snoeijs and Balashova 1998; Witkowski et al. 2000). Lithostratigraphy and diatom data were compiled using Tilia v.1.7.16. software (Grimm 2011) and CoreDRAW.

The chronology of the Lake Lilaste sediment sequence is based on six AMS (Poznan Radiocarbon Laboratory, laboratory code: Poz) and two conventional ¹⁴C dates (Institute of Geology, Tallinn University of Technology, laboratory code: Tln) described in Table 1. An age-depth model (Fig. 2) was produced using the IntCal13 calibration dataset (Reimer et al. 2013) and the OxCal 4.2.4 deposition model (Bronk Ramsey 2009, 2013), where ¹⁴C dates were combined with lithological boundaries; additionally, spheroidal fly ash particles (SFAP) were used to date upper loose sediments. The peak in SFAP at core depth 218 cm is at the model year 1982 \pm 10 AD,

which corresponds to the maximum air pollution according to the Latvenergo AS report on emissions in the atmosphere. In the current study, the radiocarbon ages were calibrated at a 95.4% probability range, and weighted averages before present were used (cal yr BP, 0=1950 AD).

Results

Lithostratigraphy and chronology

The lithostratigraphy of the Lake Lilaste core is presented in electronic supplementary material 1, and the sediment sequence is divided into 10 lithostratigraphical units (Fig. 3). Basal sediments consist of clayey silt with sand (Li-1; 1245–1260 cm) and a distinct sand layer (Li-2; 1226–1245 cm), where a piece of wood was dated at 11,160±60 cal yr BP (Table 1). Above the minerogenic sediments lies detritic gyttja (Li-3; 1218–1226 cm) that is rich in plant remains, and a piece of wood from this layer was dated at 10,890±80 cal yr BP (Table 1). Lacustrine lime (Li-4; 1193–1218 cm) is replaced by silty gyttja (Li-5; 1130–1188 cm) that consists of coarse detritus gyttja (5 cm thick), calcareous gyttja (20 cm thick), and silty gyttja layers (38 cm thick). There is a distinct boundary at 1168 cm (Li-5), where calcareous gyttja with a carbonate content of 15–20% is replaced by silty gyttja in which the carbonate content decreases to 5%. Most of the sequence contains silty gyttja with variable OM content. OM reaches its maximum of 50–60% in a gyttja layer (Li-7; 980–1060 cm), where the bulk gyttja is dated at 8600±40 cal yr BP. In silty gyttja (Li-8; 540–980 cm), constant fluctuation in OM (14–46%) and MM content (42–82%) is observed. A distinct layer with a sharp boundary and gradual colour change at 522–540 cm depth (Li-9) was deposited between 4520±100 and 4270±80 cal yr BP. In the upper part (Li-10; 200–522 cm) OM content gradually increases and is stabilized at approximately 42%, except for the interval 225–270 cm where a minor decrease to 35% in OM content is observed. MS shows high values of $9\text{--}16 \times 10^{-5}$ SI in clayey silt and decreases in sand. Once more, higher values ($20\text{--}21 \times 10^{-5}$ SI) are observed in silty gyttja (Li-5). At 1060–1130 cm depth (Li-6), MS decreases from 15 to 4×10^{-5} SI, and upwards, MS value fluctuates approximately 0.4×10^{-5} SI.

Diatom stratigraphy

A total of 253 diatom taxa representing 76 genera were identified. The most common diatom species of the Lake Lilaste sequence are displayed in Fig. 4a. Table describing the diatoms in detail is given in electronic supplementary material 2. The diatom stratigraphy was split into nine diatom assemblage zones (DAZ) using CONISS.

The basal clayey silt and sand were barren of diatoms. The first diatoms, mostly planktonic freshwater *Aulacoseira ambigua*, *A. granulata* and *Stephanodiscus parvus*, and halophilous *Cyclotella meneghiniana*, appear in detritic gyttja at 10,730±90–10,950±90 cal yr BP (DAZ-1). Subsequently, in carbonaceous sediments and silty gyttja up to 1100 cm, diatoms are not preserved. Mass occurrence of *Fragilaria* spp., represented by *Staurosira construens* and *S. venter*, occurs in silty gyttja at 1090 cm (DAZ-2). At 1065 cm, *Fragilaria* spp. abruptly declines, whereas planktonic freshwater flora, including *A. ambigua*, *A. granulata*, *Cyclostephanos dubius*, *Cyclotella comta*, *Stephanodiscus parvus*, expands (DAZ-3). Sporadically appearing at the same time are the first marine/brackish-water taxa, such as *Navicula cincta* and *Planothidium delicatulum*, and indifferent taxa including *Amphora libyca*, *A. pediculus*, *Cocconeis placentula*, and *C. neothumensis*. In the lowest part of DAZ-4 *Stephanodiscus parvus* dominates and thereafter is gradually replaced by *A. ambigua* and *A. granulata*. Two peaks of diatom flora that prefer brackish or high-conductivity water are observed at ca. 7600±40 and ca. 7300±40 cal yr BP. Significant abundances of marine/brackish (up to 23%) and halophilous (up to 27%) diatoms are observed in DAZ-5 and DAZ-6. Marine/brackish taxa primarily are represented by planktonic *Cyclotella*

choctawhatcheeana and *Fragilariopsis cylindrus*, and periphytic *Achnanthes fogedii*, *P. delicatulum*, *P. lemmermannii*, *Cocconeis hauniensis*, *Navicula gregaria*, *N. perminuta* and *N. rhynchocephala*, and the main component of the halophilous taxa is *C. meneghiniana*. Small-sized *Fragilaria* spp., with a brackish-water affinity, are mostly represented by *Fragilaria sopotensis* (peak 51%), *Opephora mutabilis* and *Staurosira punctiformis*. A characteristic feature in DAZ-5 is the amount of *Chaetoceros* spp. resting spores, which fluctuates between 5 and $14 \times 10^5 \text{ cm}^{-2} \text{ yr}^{-1}$. The diatom community of DAZ-7 is dominated by planktonic freshwater taxa, representing 75–90% of the total. In the upper part of the sequence (DAZ-8, 9), freshwater planktonic taxa decrease in abundance and are gradually replaced by small-sized fragilarioid taxa.

Discussion

The Coastal Lowland and Lake Lilaste areas were deglaciated at approximately 14,000 cal yr BP (Vassiljev and Saarse 2013) and were flooded by the BIL, the coastal formations of which are located at 22–26 m a.s.l. (Grinbergs 1957). Modelling results (Vassiljev and Saarse 2013) suggest that the water level reached up to 16 m a.s.l. at 11,700 cal yr BP (Fig. 5). The rapid drainage of the BIL at Mt. Billingen through the southcentral Swedish lowlands at 11,690±10 varve yr BP (Andrén et al. 2002) or 11,653±99 cal yr BP (Rasmussen et al. 2006) lowered the water level by approximately 25 m to the ocean level. In the Baltic Sea basin, the Yoldia Sea was established, and the water level at Lake Lilaste was approximately 9 m b.s.l. (Fig. 5) so that the lake water depth was approximately 3 meters. Clayey silt (Li-1) at the basal part of the sequence, most likely deposited in the BIL and sand (Li-2), might have been deposited during drainage of the BIL or erosion at the time of the Yoldia Sea. Two AMS ^{14}C dated wood remains from sand (Li-2) and thin gyttja (Li-3) layers of Lake Lilaste ca. 11,160±60 and ca. 10,890±80 cal yr BP, respectively, indicate the possibility that these sediments accumulated during the Yoldia Sea phase. Wood remains suggest that a low water level existed in the area during the Yoldia Sea phase. Modelling results (Saarse et al. 2003) show that the water level in the Gulf of Riga during the Ancyclus Lake transgression (10,300 cal yr BP) was about -14 m a.s.l., so that the Yoldia Sea was lower than that (Fig. 5). In the depression of Lilaste, a shallow lake was formed. Sands around the Lilaste depression were the source of material forming the surrounding dunes (Fig. 1b) during the Early Holocene when the pristine newly emerged landscape was covered by sparse vegetation. Numerous plant macroremains and diatoms in the gyttja layer (Li-3) indicate a shallow eutrophic freshwater lake environment, and suggest that the Lilaste basin was not connected with the Baltic Sea basin at the end of Yoldia Sea phase. The existence of a eutrophic and high-conductivity state is supported by the presence of the small planktonic diatoms *Stephanodiscus parvus* and *Cyclotella meneghiniana* (Anderson 1990; Witak 2013), as well as meroplanktonic *Aulacoseira ambigua* and *A. granulata*, indicating a comparatively shallow, nutrient-rich environment (Shear et al. 1976; Kilham 1990). Intense nutrient input was caused by rivers that collected water from sparsely vegetated catchment areas that favoured nutrient supply to the lake (Seppä et al. 2000).

No diatoms were preserved in lacustrine lime (Li-4) and silty gyttja (Li-5). Therefore, the sedimentation environment can be described only on the basis of lithostratigraphical evidence. According to the aforementioned diatoms in the detritic gyttja (Li-3, Fig. 3), which is full of plant macroremains, a shallow freshwater environment is indicated. The water level gradually began to rise (Fig. 5), and due to the carbonate-rich till that lies close to the eastern shore, lacustrine lime precipitated and was replaced by an accumulation of silty gyttja (Li-5) approximately 9750±120 cal yr BP. Most likely, the Ancyclus Lake transgression ca. 10,300 cal yr BP (Saarse et al. 2003) induced a groundwater table rise in coastal areas, which successively caused an increase in lake water depth.

An increase in MM content along with mass occurrence of small-sized fragilarioid taxa (DAZ-2; Fig. 4) indicates a period of environmental instability due to increased turbidity and higher nutrient content (Haworth 1975; Seppä and Weckström 1999; Anderson 2000) that lasted for approximately 300 years (8720±50–9000±70 cal yr BP). It is difficult to determine the possible

cause of the transition of environmental conditions in the lake water basin, as diatoms in lower sediments are not preserved. However, mass occurrence of *Fragilaria* spp. could be elucidated by fluctuations in groundwater table or increased erosion. The appearance of the marine/brackish-water taxa at 8690±50 cal yr BP (DAZ-3, Fig. 5) suggests that the water level in the Baltic Sea basin was close to the Lake Lilaste threshold. The exact position of the threshold at that time is unknown, but it was most likely higher than -10 m a.s.l. (sediment surface) and lower than -3.6 m, corresponding to the modelled water level at 7800 cal yr BP (Fig. 5). It appears that the water level rise in Lake Lilaste started simultaneously with the Litorina Sea transgression in the Baltic Sea at 8500 cal yr BP (Berglund et al. 2005).

Since 8720±50 cal yr BP (DAZ-3), diatom stratigraphy indicates freshwater conditions throughout the sediment sequence, except for the time span between 6730±40 and 4790±80 cal yr BP (DAZ-4), when the abundance of freshwater taxa substantially declines, and is replaced by marine/brackish-water, halophilous and indifferent taxa (Fig. 4b). Together with marine/brackish-water taxa, nutrient rich environment indicators such as *Stephanodiscus parvus* and *Cyclotella meneghiniana* (Heinsalu et al. 2007; Weckström and Juggins 2005) dominate. The enhanced nutrient content in Lake Lilaste could be explained by occasional mixing of brackish and fresh water that, according to Head (1976), promotes biological productivity and enrichment with organic compounds. Similar eutrophication due to marine water intrusion has been previously observed in several isolation basins (e.g., Westman and Hedenström 2002; Yu et al. 2004; Grudzinska et al. 2013, 2014).

Prior to the clear evidence of a constant presence of brackish water in Lake Lilaste due to the long lasting Litorina Sea transgression since 6730±40 cal yr BP, it is possible to refer to two significant sea water surges at ca. 7600±40 and ca. 7300±40 cal yr BP (DAZ-4) based on lithological records (peak of MM) and the presence of marine/brackish and halophilous diatoms, probably related to strong north-westerly gales that eroded coastal formations (Fig. 1b). Several factors favoured persistent sea water intrusion into isolation basins such as Lake Lilaste. The first such factor is the sea level altitude in relation to the threshold of the isolation basin. Diatom records and modelled data show that the Litorina Sea level reached close to the threshold at 6730±40 year ago. At the same time, eustatic sea level (ESL) rise significantly slowed down (Lambeck et al. 2014), suggesting that land uplift in the area must have been similar to the ESL change rate (Fig. 5). The second factor is the location of the isolation basin in relation to the position of the sea coast in the past. In the Mid-Holocene, Lilaste was much closer to the sea coast than it is currently. According to Eberhards (2003), the southern part of the Gulf of Riga is an advancing depositional coast, based on analysis of historical maps and topographic material, and the coast in the area of the mouth of the River Gauja (5 km SW from Lake Lilaste, Fig. 1b) has advanced by 400–500 m over a period of 300 years. Hence, it is considered that dunes and foredunes in an area 1 km wide between Lilaste and the sea coast formed over the last 4200 years. Finally, ice conditions in the coastal zone and ground frost conditions play significant roles in the protection of coastal formations against erosion. At the beginning of the Mid-Holocene, the mean winter temperature in the Baltic region was higher than the modern winter temperature (Veski et al. 2014). Therefore, climatic conditions along with higher salinity (Andrén et al. 2000) during the Holocene thermal maximum (HTM) were unfavourable to the formation of a permanent sea ice cover. The highest storm surge level at the southern end of the Gulf of Riga at Daugavgriva reached up to +2.14 m a.s.l. during the 1969 hurricane (Eberhards 2003). Consequently, due to the absence of sea ice and the amelioration of ground frost, severe autumn and winter storm surges washed away newly formed foredunes, and sea water freely entered the lake. If the Litorina Sea level was at the same altitude as the Lilaste threshold, strong north-westerly winds could cause sea level rise and brackish water flow into the lake.

The continuous Litorina Sea water intrusion confirms the increase in the abundance of marine/brackish-water and halophilous diatoms, and this is also supported by the presence of typical Litorina Sea species such as *Achnanthes fagedii* (Snoeijs and Kasperovičienė 1996; Witkowski et al. 2000). The extended influx of brackish water into the basin is indicated by the diatom

assemblage. Additionally, *Chaetoceros* resting spores and some ebridian remains in DAZ-4 evidence changes in environment. *Chaetoceros* spp. valves of vegetative cells are weakly silicified, and thus usually dissolve in the water column, as fossil diatoms occur very rarely (Witak et al. 2011) but their spores are quite often preserved in sediments. *Chaetoceros* resting spores are formed to survive environmental stress, such as nutrient depletion or changes in salinity of the freshwater environment. Furthermore, in the studies of Andrén et al. (2000) and Witak et al. (2011), it was concluded that the appearance of *Chaetoceros* resting spores in sediments is linked to an inflow of saline waters rich in nutrients, and the highest abundance of spores was observed in Litorina Sea sediments. In addition, increased access to nutrients or more brackish conditions supports the presence of *Ebria tripartita* (Andrén et al. 2000).

The most intense marine water influence on the Lilaste basin was observed from 6150±30 to 5500±40 cal yr BP. During this time, planktonic freshwater diatoms, such as *Aulacoseira ambigua*, *A. granulata*, and *Cyclotella comta* almost disappear, and only *Stephanodiscus parvus*, which occurs in a nutrient rich environment with high conductivity, reaches its maximum. The dominance of diatom taxa with relatively small proportions leads to the assumption of increased silica utilization due to enhanced nutrient loading caused by brackish water inflow. Limited amount of silica in the environment promotes small-sized diatoms, such as *S. parvus* and small-sized *Fragilaria* spp., to build frustules quicker and outcompete large-sized diatoms. Although silica utilization was rather high, frequent brackish water intrusions and the inlet of river waters provided additional nutrients required for the formation of diatom frustules.

The distinct brackish water intrusion from 6700±40 to 4800±80 cal yr BP coincides with the stillstand in areas in Norway with low land uplift rates at 6500–4900 cal yr BP (Balascio et al. 2011). Brackish water influence on the lake was interrupted at ca. 4800±80 cal yr BP. An increase in OM and the dominance of freshwater diatoms indicates stabilization of the freshwater environment for ca. 300 years. In the time interval 4500±100–4200±80 cal yr BP, a rapid increase in MM and a reappearance of brackish water diatoms were observed. The recurrent marine water inflows coincide with the second Litorina Sea transgression described by Veinbergs (1979). However, the rate of ESL rise decreased from 1 to 0.5 mm yr⁻¹ around that time (Fig. 5; Lambeck et al. 2014) and does not support transgression. Increased storm frequency resulting in erosion of the coastal formations could be one reason why brackish water diatoms reappear.

The lithology of the Lake Lilaste sequence sustains the hypothesis regarding long lasting, intermittent sea water inflow for approximately 4500 years. Along with the appearance of marine/brackish-water diatoms since 8720±50 cal yr BP, distinct fluctuations in OM and MM contents are observed. A gradual OM increase during the isolation, confirming the formation of a stable freshwater environment in the lake, has been observed in other studies (Saarse et al. 2009; Long et al. 2011; Grudzinska et al. 2013; 2014).

In accordance with observations in Estonia and Finland, the diatom compositions do not support the idea of the twofold Litorina Sea transgression suggested earlier, but support one main transgression event. However, minor fluctuations are normal, especially during strong north-westerly winds or storms. During the Late Holocene, active longshore sand transport and additional mineral matter from rivers contributed to bars, foredunes, spits and wider beach formations. Presently, the height of foredunes in the study area reaches up to 3–6 m a.s.l., but inner dunes up to 20 m a.s.l. high (Fig 1b) protect the current lake from storm surges.

According to simulations, the basin finally became isolated as a separated lake without any brackish-water intrusion at ca. 4200 cal yr BP. The isolation of Lake Lilaste promoted the reduction of nutrient loads, and recovery was triggered for the appearance of *Aulacoseira islandica* and *A. subarctica* (Gibson et al. 2003). Some brackish-water diatoms identified in the upper part of the sequence could have originated from reworked material. Anthropogenic influence on the lake ecosystem is indicated by increases in MM content and small-sized fragilarioid taxa, along with a decline of freshwater planktonic taxa at 400±50 cal yr BP.

Conclusions

Lake Lilaste is characterized by a long and variable sediment sequence beginning at 11,400 cal yr BP and starting from clayey silt with sand interbeds, which deposited during the Yoldia Sea phase, and the basal clayey silt could have deposited in the BIL.

During the Yoldia Sea, Lilaste existed as an isolated water body. Before the Ancyclus Lake transgression, the water level in Lilaste should have decreased below the threshold deduced from deposition of detritus gyttja and lacustrine lime, which is unfortunately barren of diatoms.

In the Ancyclus Lake phase, sedimentation of lacustrine lime occurred in the freshwater isolated basin.

An occasional influx of sea water to the lake started concurrently with the Litorina Sea transgression ca. 8700 cal BP, but the diatom composition indicates preservation of freshwater conditions up to ca. 6700 cal BP.

The diatom composition and interpolated radiocarbon dates suggest that brackish water environments lasted ca. 2200 years in two time periods 6700±40–4800±80 and 4500±100–4200±80 cal yr BP, caused by the Litorina Sea transgression. During this transgression, the sediment composition became more minerogenic.

Long-term influxes of brackish water into Lilaste could be explained by its proximity to the sea, which facilitated seawater inflow during strong north-westerly winds and storms, and the persistence of a brackish habitat for diatoms.

Diatom composition and sediment lithostratigraphy indicate that Lake Lilaste was isolated at two times: first during the Yoldia and Ancyclus phase, and second during the Limnea Sea.

Due to a very small land uplift rate close to zero (Ekman 1996), the Litorina Sea transgression occurred almost 1000 years later than in lakes at high isolines, as in Estonia and Finland.

Acknowledgements

The authors are thankful to A. Heinsalu and an anonymous referee for useful comments and suggestions for improving the manuscript. The Elsevier language editing staff is acknowledged for correcting the English. The study was supported by ESF Grant 9031, IUT 1-8, and Doctoral Studies and Internationalisation Programme DoRa.

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Table 1 Radiocarbon ^{14}C dates from the Lake Lilaste sequence

Depth from the water surface (cm)	Age (^{14}C BP)	Calibrated age at 95.4% (weighted average) (cal yr BP)	Lab. No.	Material
380	2385±30	2440–2340 (2380±30)	Poz-63856	bulk gyttja
515–520	3793±60	4410–4090 (4210±80)	Tln-3500	bulk gyttja
540–545	3920±55	4800–4370 (4520±100)	Tln-3497	bulk gyttja
660	4880±30	5650–5490 (5610±30)	Poz-63857	bulk gyttja
878	6370±40	7420–7240 (7300±40)	Poz-57601	bulk gyttja
1050	7810±35	8700–8520 (8600±40)	Poz-63858	bulk gyttja
1222	9630±60	11190–10730 (10890±80)	Poz-48438	wood
1240	9710±50	11,240–11,070 (11160±60)	Poz-48439	wood

Figure captions

Fig. 1 **A** Location of the study area shown on the overview map. Dashed lines show apparent uplift in mm yr^{-1} (Ekman 1996). **B** Modern topography in the surroundings of Lake Lilaste, with an indication of the sampling site

Fig. 2 Age-depth model based on the OxCal deposition model at a 95.4% probability range

Fig. 3 Loss-on-ignition, accumulation rate and magnetic susceptibility from the Lake Lilaste core

Fig. 4 **A** Percentage diagram of selected diatom taxa from Lake Lilaste (arranged according to their salinity tolerance and habitat classification), diatom accumulation rate and distribution of *Chaetoceros* resting spores (1–Planktonic marine/brackish taxa; 2–Periphytic marine/brackish taxa; 3–Planktonic halophilous taxa; 4–Periphytic halophilous taxa; 5–Periphytic indifferent taxa; 6–Small-sized fragilarioid taxa with brackish-water affinity; 7–Small-sized fragilarioid taxa; 8–Planktonic freshwater taxa; 9–Periphytic freshwater taxa). **B** Percentage summary diagram of the diatom assemblage from Lake Lilaste

Fig. 5 Water level curve of Lake Lilaste, where Baltic Sea stages, lithological units (Li) and diatom assemblage zones (DAZ) are shown. Eustatic sea level data are based on Lambeck et al. 2014

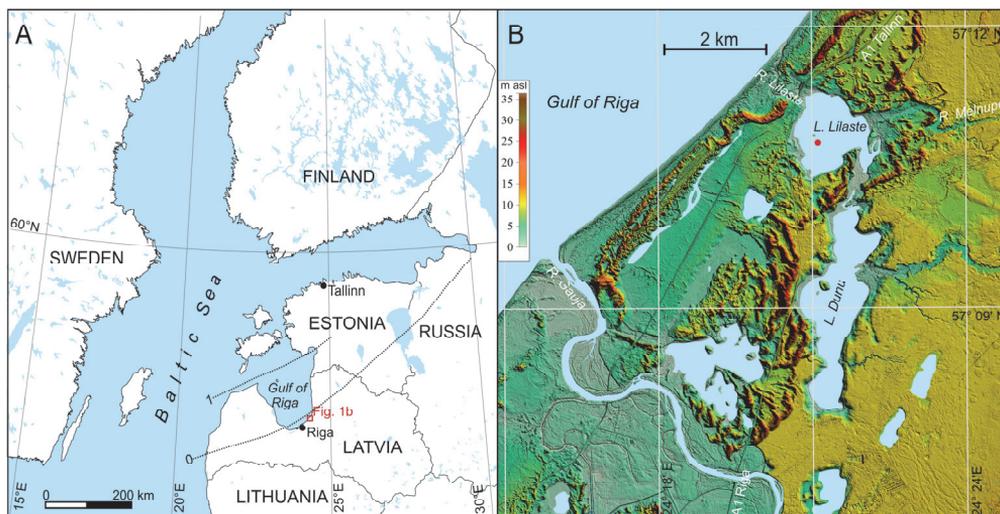


Fig. 1

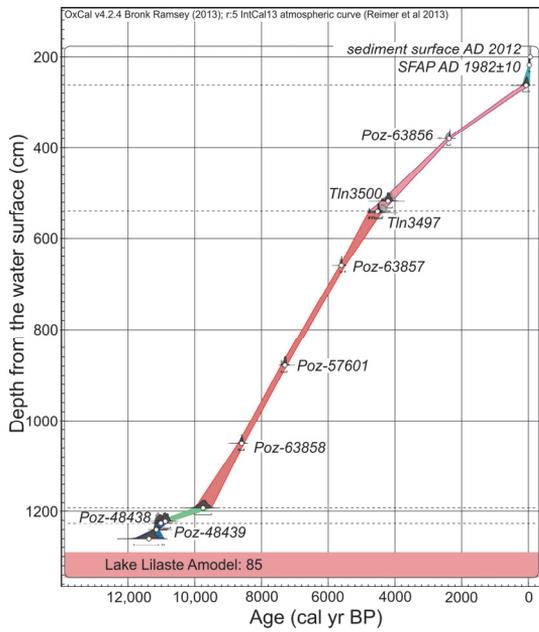


Fig. 2

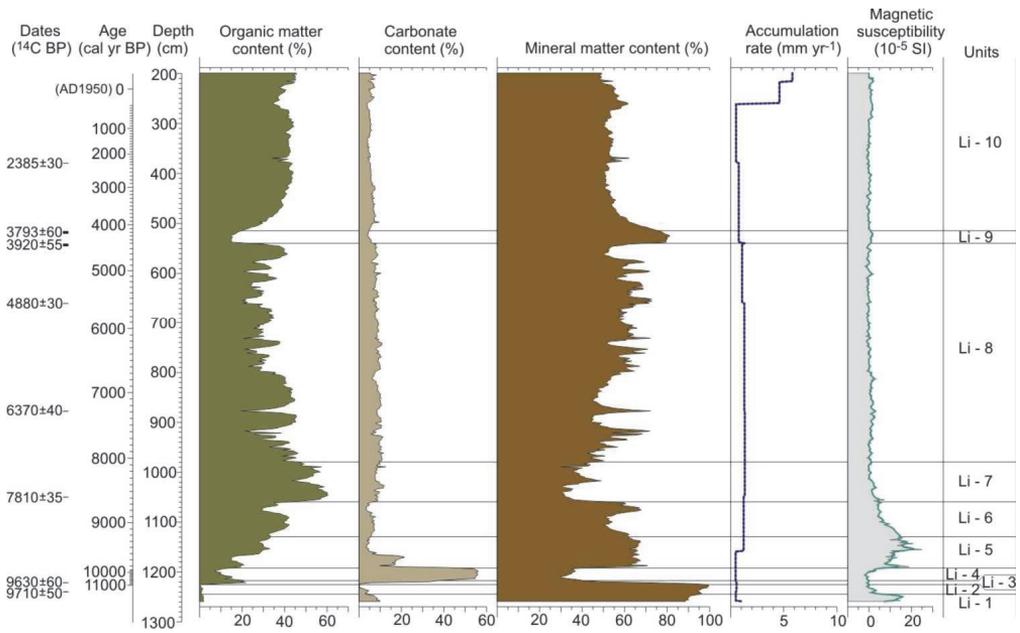


Fig. 3

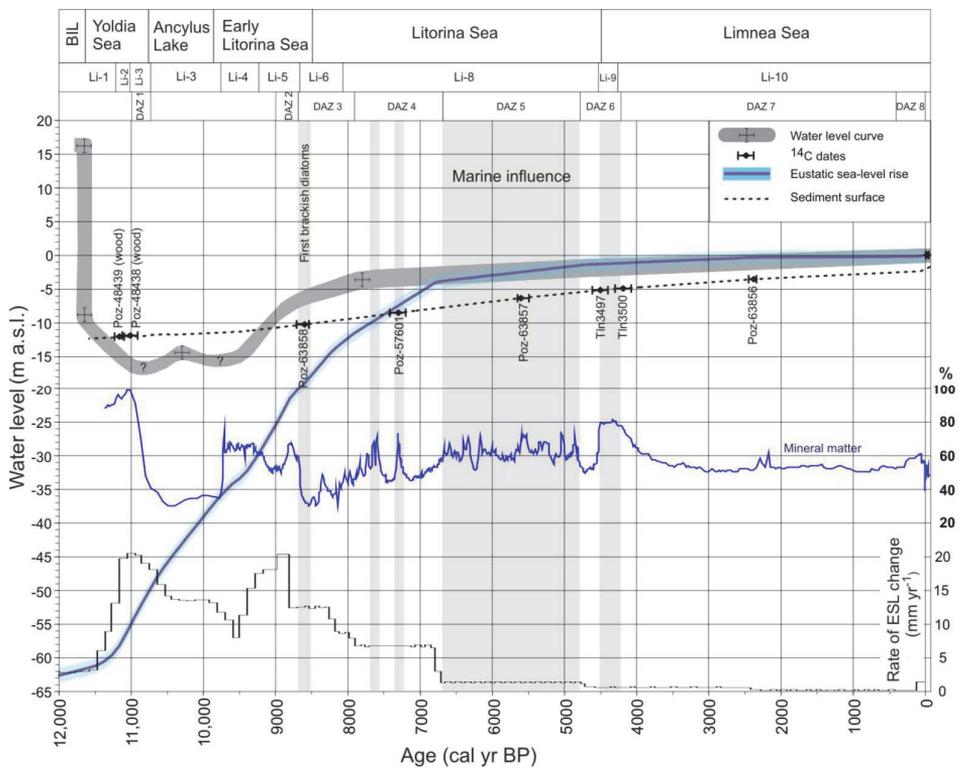


Fig. 5

Electronic Supplementary Material 1. Lithology of the studied Lake Lilaste sequence. OM—organic matter, MM—mineral matter

Depth from the water surface (cm)	Calibrated age, weighted average (cal yr BP)	Lithological unit	Description
200–522	Present–4270±80	Li-10	Silty gyttja Colour: brown, homogeneous. MM gradually decreases to 53%. At depth 270 cm MM shortly increases to 60–62% and at upper part it decreases again. Carbonate content is low, ca. 5–6%. OM slowly increase to 40–42%.
522–540	4270±80–4520±100	Li-9	Silty gyttja Gradual colour change from grey to brown. Stable high MM content ca. 80–81% and low OM content ca. 15%.
540–980	4520±100–8070±40	Li-8	Silty gyttja with sand interlayers Colour: dark brown, at 595–800 cm laminated. MM with distinct fluctuations increases from 42 to 74%, two peaks at depth of 878–880 cm and 915–920 cm are observed. OM content varies between 18 and 40%, carbonate content – ca. 8%. At 595–800 cm sand interlayers are observed.
980–1060	8070±40–8680±50	Li-7	Gyttja Colour: black/brown. MM is around 37%, the maximum peak (48%) is at depth 1016–1020 cm. OM reaches its maximum 60%. Carbonate content fluctuates around 8.5–9%.
1060–1130	8680±50–9240±90	Li-6	Silty gyttja Colour: black/brown. In the upper part of unit MM rises up to 68%, while OM decreases from 43 to 29%, carbonate content is ca. 6%.
1130–1193	9240±90–9750±120	Li-5	Silty gyttja Unit contains three layers: coarse detritus gyttja (1188–1193 cm), calcareous gyttja (1168–1188 cm) and silty gyttja (1130–1168 cm), colour of sediments changes from dark grey at the bottom to light yellowish brown in the middle, and upwards turns light greyish brown. MM content is around 65%, OM increases up to 30% and carbonate content at the bottom fluctuates around 21% and upwards decreases up to 5%. The unit is characteristic by high MS values that increases from 4 up to 21×10^{-5} SI.
1193–1218	9750±120–10,730±90	Li-4	Lacustrine lime Colour: light beige. Carbonate content fluctuates around 54–56%. MM fluctuates around 30–38%, and OM content decreases from 17 to 12%.

1218–1226	10,730±90– 11,020±90	Li-3	Detritic gyttja with plant remains Colour: dark greyish brown. OM increases till 20%, while MM decreases up to 40.5%. MS values around 0.
1226–1245	11,020±90– 11,210±90	Li-2	Sand Colour change from brownish grey to yellow. Fine sand with low MS values, high MM content 95–99.5%, very low OM and carbonate content.
1245–1260	11,210±90– 11,375±190	Li-1	Clayey silt with sand Colour: reddish brown. Low OM and carbonate content, high MM content 88–95%, MS maximum value is 16.5×10^{-5} SI.

Electronic Supplementary Material 2. Diatom stratigraphy of the Lake Lilaste sequence

Depth from the water surface (cm)	Calibrated age, weighted average (AD / cal yr BP)	DAZ	Description
200–225	Present to AD 1960±15	9	Small-sized freshwater fragilarioid taxa dominate (ca. 50%) <i>Staurosira construens</i> , <i>S. venter</i> , <i>S. construens</i> var. <i>binodis</i> . They are accompanied by freshwater taxa, including planktonic <i>Aulacoseira ambigua</i> and <i>A. granulata</i> , and periphytic <i>Geissleria schoenfeldii</i> , <i>Planothidium frequentissimum</i> , <i>P. joursacense</i> . At the zone bottom <i>Fragilaria</i> spp. with brackish-water affinity increase and abundance of indifferent flora <i>Amphora pediculus</i> and <i>Cocconeis neothumensis</i> is observed.
225–280	AD 1960±15–420±50	8	Planktonic freshwater taxa decreases from 74% to 60%, whereas small-sized fragilarioid and indifferent taxa gradually increases from 20 to 30% and 1 to 6%, respectively. <i>Aulacoseira granulata</i> is consistently ca. 11% and <i>Cyclostephanos dubius</i> ca. 3%, while <i>A. ambigua</i> decreases from 58 to 38%.
280–518	420±50–4220±80	7	Maximal abundance of planktonic freshwater diatoms up to 84% – <i>Aulacoseira ambigua</i> (ca. 60%). Periphytic marine/brackish-water <i>Planothidium delicatulum</i> is observed sporadically. Small-sized fragilarioid taxa occurs rather rarely, being represented most often by <i>Pseudostaurosira brevistriata</i> , <i>Staurosira construens</i> and <i>S. venter</i> . DAR decreases upwards from 3 to 0.1×10^6 valves $\text{cm}^{-2} \text{yr}^{-1}$.
518–570	4220±80–4790±80	6	At the bottom of zone freshwater flora dominates, but at depth 542 cm sharply drops from 90 to ca 50%. At the same time abundance of marine/brackish-water, halophilous, indifferent, small-sized fragilarioid taxa rises. Marine/brackish taxa are represented by planktonic <i>Cyclotella choctawhatcheeana</i> , periphytic <i>Planothidium delicatulum</i> and <i>P. lemmermannii</i> . Increase up to 20% experiences planktonic halophilous <i>Cyclotella meneghiniana</i> . Together with marine/brackish-water and halophilous taxa freshwater planktonic <i>Cyclotella atomus</i> appears and increase of <i>Cyclostephanos dubius</i> and <i>Stephanodiscus parvus</i> is observed. DAR rises up to 9×10^6 $\text{cm}^{-2} \text{yr}^{-1}$ along with increase of marine/brackish and halophilous taxa.
570–800	4790±80–6730±40	5	A distinct increase in the abundance of marine/brackish-water, halophilous, indifferent and small-sized fragilarioid with brackish-water affinity taxa. Halophilous taxa were primarily represented by planktonic <i>Cyclotella meneghiniana</i> , and the main components of marine/brackish-water taxa are <i>Cyclotella choctawhatcheeana</i> , <i>Planothidium delicatulum</i> , <i>P. lemmermannii</i> , <i>Achnanthes fagedii</i> and <i>Navicula gregaria</i> . In the middle of zone at depth 650–730 cm planktonic freshwater diatoms almost disappear, except

Stephanodiscus parvus that reaches its maximum (49%). Singular peak of *Fragilaria sopotensis* (51%) at 606 cm and distinct peak of planktonic marine/brackish-water *Fragilariopsis cylindrus* (11%) at 762 cm is observed. Noteworthy characteristic feature is accumulation rate of *Chaetoceros* resting spores that fluctuates between 5 and $14 \times 10^5 \text{ cm}^{-2} \text{ yr}^{-1}$.

800–960	6730±40– 7920±40	4	Freshwater planktonic <i>Aulacoseira granulata</i> , <i>A. ambigua</i> , <i>Cyclotella dubius</i> , <i>Stephanodiscus neostreae</i> and <i>S. parvus</i> dominate (40%) in lower part of the zone and decrease upwards. <i>Cyclotella ocellata</i> similarly to <i>S. parvus</i> reaches its maximum 5% at the lower part of zone and decreases upwards. Amount of periphytic freshwater <i>Navicula scutelloides</i> increases at depth 850–910 cm. Significant rise of marine/brackish-water and small-sized fragilarioid taxa with brackish-water affinity and decline of freshwater taxa are observed at two levels 915–930 cm and 875–880 cm. At depth 878 cm appears <i>Chaetoceros</i> resting spores.
960–1065	7920±40– 8720±50	3	Dominates freshwater planktonic <i>Aulacoseira ambigua</i> which abruptly decreases at 1000–1020 cm, while increase of <i>Stephanodiscus parvus</i> (up to 10 %), <i>Cyclotella comta</i> (up to 6 %) and <i>Staurosira venter</i> (up to 8 %) is observed. <i>Aulacoseira granulata</i> increases gradually to 20 %. First appearance of marine/brackish-water diatoms, e.g., <i>Navicula cincta</i> , <i>N. rhynchocephala</i> and <i>Planothidium delicatulum</i> , and indifferent flora, such as, <i>Amphora libyca</i> , <i>A. pediculus</i> , <i>Cocconeis placentula</i> , <i>C. neothumensis</i> . DAR fluctuates from 4 to $8 \times 10^6 \text{ valves cm}^{-2} \text{ yr}^{-1}$.
1065–1100	8720±50– 9000±70	2	High abundance of small-sized fragilarioid taxa 89–91% and high DAR $15\text{--}17 \times 10^6 \text{ cm}^{-2} \text{ yr}^{-1}$. From <i>Fragilaria</i> spp. dominate <i>Staurosira construens</i> (maximum 70 %) and <i>Staurosira venter</i> (12–13%). From other freshwater diatoms periphytic <i>Geissleria schoenfeldii</i> occurs.
1218–1224	10,730±90– 10,950±90	1	Planktonic freshwater taxa dominate, such as, <i>Aulacoseira ambigua</i> (44%), <i>A. granulata</i> (13%) and <i>Stephanodiscus parvus</i> (6%), accompanied by small-sized fragilarioid taxa and planktonic halophilous <i>Cyclotella meneghiniana</i> that prefers water with high conductivity.

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