

An acoustic view into Holocene palaeoseismicity offshore southwestern Finland, Baltic Sea

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Abstract

The Baltic Sea basin consists of Palaeoproterozoic crystalline bedrock, which is to the greatest extent covered by younger Proterozoic and Phanerozoic sedimentary bedrock. The basin contains several ancient deep-seated tectonic lineaments and fracture zones that divide the seabed into blocks. High-resolution, low frequency, echo-sounding was used to study the Holocene submarine sedimentary structures and palaeoseismic activity in the northern Baltic Sea with the aim of tidying previous, more site-specific results (at Olkiluoto) into a larger regional context. Echo-sounding profiles of Holocene submarine sediments show slides and slumps, normal faults, debris flows and turbidite-type structures. The profiles show also pockmarks and other structures related to gas or groundwater seepages. Some of the observed structures are suggested to be associated with palaeoseismic activity 10,700 cal yr BP (~9500 radiocarbon years BP) in the northern Baltic Sea. Post-glacial fault activation and possible gas and/or fluid discharges under high hydraulic heads have relevance in the safety assessment of a planned spent nuclear fuel repository in the region. Evidence of post-glacial faulting in the study area is found only along old bedrock fracture zones indicating reactivation of the old fracture zones. Since no younger or repeated traces of possible seismic events were found, it corroborates the suggestion that the major seismic activity occurred within a short time during and after the last deglaciation.

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1. Introduction

High-resolution echo-sounding at low frequencies is an effective method to study sedimentary structures of Holocene marine and lacustrine sediments. Some structures, like soft sediment faults, slide/slump structures, pockmarks and debris flows have been connected with palaeoseismicity in the post-glacial period (e.g. Laberg and Vorren, 2000; Kotilainen and Hutri, 2004). In

the Olkiluoto area, in the Gulf of Bothnia (Fig. 1), two NW–SE trending bedrock fracture zones are suggested to have been reactivated after the last Weichselian deglaciation and to have triggered several faults and other “disturbance structures” observed in the Holocene sediments (Kotilainen and Hutri, 2004). The age estimation for the event(s) is 10,650–10,200 cal yr BP (Hutri et al., 2007). Post-glacial rebound and ridge push from the Mid-Atlantic have been suggested as causative mechanism for post-glacial faulting in Fennoscandia. A bedrock modelling study of deglaciation scenarios in the Olkiluoto area (Hutri and Antikainen, 2002) supported the concept of post-glacial faulting connected to glacial

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rebound along NW–SE and SW–NE trending fracture zones in the area. These phenomena have immediate relevance for the planned repository of spent nuclear fuel in the area. The aim of this study is to examine the previously obtained results in a larger context, within a radius of ca. 200 km of the Olkiluoto site (Fig. 1). The major tectonic zones and lineaments of the Baltic region (Fig. 1) provide a basis for the seismotectonical correlation (Flodén, 1982; Koistinen et al., 1996, 2001; Korja and Heikkinen, 2005, and lineament interpretations by Kuivamäki, 2005).

2. Study area

The study areas are located in the southern Bothnian Sea, the Archipelago Sea, and the northern Baltic Proper (Figs. 1 and 2), within a distance of 200 km of Olkiluoto. These are the only areas with available modern digital echo-sounding and reflection seismic data in the vicinity of Olkiluoto. Although a large number of old acoustic-seismic profiles exist in paper form, examination of these would not have provided any detailed information due to poor resolution. They cover all together $\sim 1200 \text{ km}^2$ (2200 profile km) (Fig. 2).

The current stress field in Fennoscandia is dominated by the tectonic forces sustained by the Mid-Atlantic Ridge push. The NW–SE direction of tectonic compression is dominant (Scandinavian stress map, http://www-gpi.physik.uni-karlsruhe.de/wsm/home/index_noflash.html) (Fig. 1). However, the maximum horizontal stress also depends on depth, and the regional stress fields may consist of components of tectonic and local effects (e.g. glacial rebound) (Martin et al., 1990). During and just after the latest deglaciation the whole study area was submerged more than 150 m (Eronen et al., 2001).

In the study area, the present land uplift varies ~ 3 –6 mm/a (Ekman and Mäkinen, 1996), from south to north. Several NW–SE and NE–SW trending fracture zones and lineaments are typical for the whole study area. A lineament interpretation of the area after Kuivamäki (2005) is presented in Fig. 2. The long, parallel NW–SE running faulted fracture zones in the Bothnian Sea belong to the Hämösand–Pori fault zone (Figs. 1 and 2) (Flodén, 1982; Koistinen et al., 1996, 2001).

The present seismicity in the study area is low, and according to the Fennoscandian earthquake database (<http://www.seismo.helsinki.fi/bul/index.html>) only

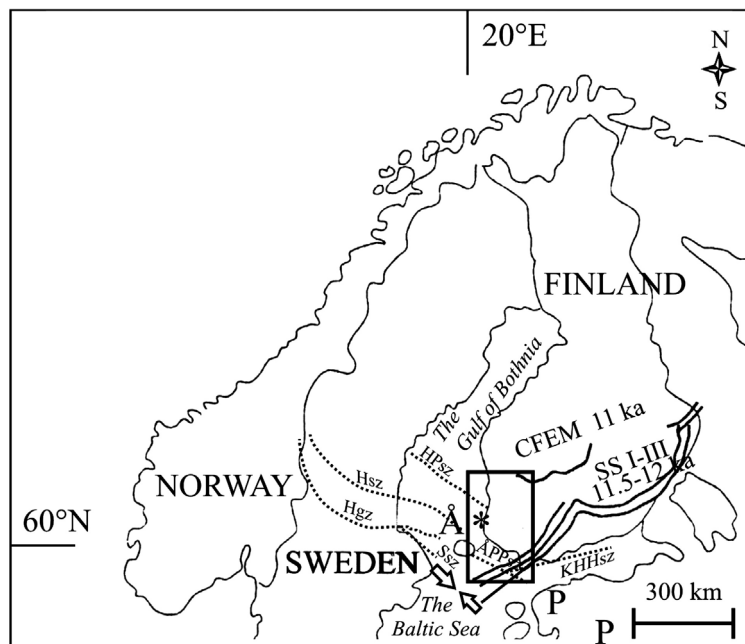


Fig. 1. Some major tectonic zones in the northern Baltic Sea region according to Korja and Heikkinen (2005), Koistinen et al. (1996, 2001) and Flodén (1982): Hsz = Hassela shear zone, Hgz = Hagsta deformation zone, Ssz = Singö shear zone, ÄPPsz = Åland–Paldiski–Pskov shear zone, HPsz = Hämösand–Pori shear zone, KHHsz = Kökar–Hanko–Helsinki shear zone. The study area is marked with a rectangle and the letters SS I–III refer to the Salpausselkä ice margin formations and CFEM to the Central Finland End Moraine formation. The regional stress field is marked with arrows. The Olkiluoto area with previous observations of Holocene palaeoseismicity is marked with a star in the figure. Letters Ä, P and P refer to the Åland–Paldiski–Pskov seismic zone (e.g. Saari, 1998a, 2000).

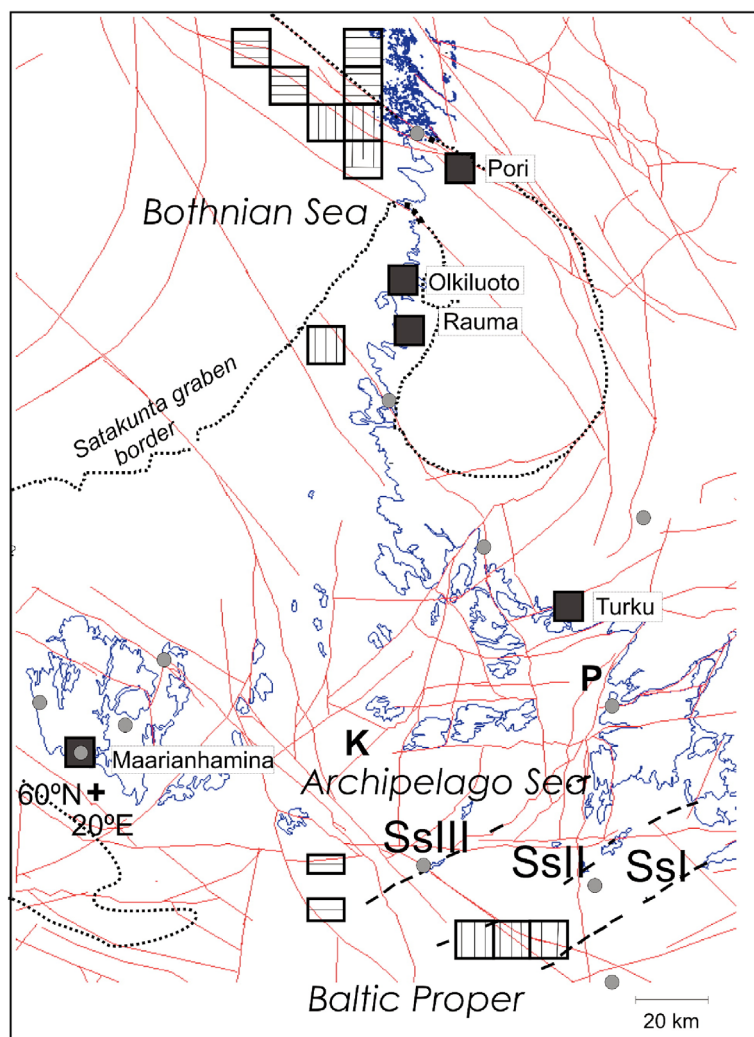


Fig. 2. The study areas are indicated on the map with squares according to the Finnish map sheet division. The locations of the survey profiles are marked as horizontal or vertical lines inside the map sheet squares, meaning east–west or north–south directions. The profiles are located about 500 m from each other. The bedrock fracture zones according to Kuivamäki (2005) are shown with red lines. The tectonic zones refer to Flodén (1982) and Koistinen et al. (1996, 2001). The letters Ss I–Ss III refer to the Salpausselkä ice margin formations. Letters K and P represent Kråkskär and Paimionlahti areas referred in the text. The earthquake epicentres (1375–1995) with magnitude $M \geq 1.5$ in the area (Saari, 1998b) are marked with grey dots.

approximately ten small ($M < 3$) earthquakes have occurred in the area since 1375. Most of these have been located along the NW–SE trending fracture zones (Fig. 2). The 600 km long and 150 km wide Åland–Paldiski–Pskov (Fig. 1) zone has a somewhat higher seismicity (Saari, 1998a,b). Small earthquakes may occur in the lengthening of quietly creeping fractures reflecting ongoing bedrock deformation processes over a time span of thousands of years (Saari, 1998a). Recent GPS measurements carried out in the Olkiluoto area (Ollikainen et al., 2004) have affirmed very small crustal movements indicating the motion of the Eurasian plate.

The bedrock of the Baltic Basin consists of Svecofennian crystalline rocks on top of which sedimentary rocks of different ages have been deposited. The only known sedimentary rock unit in the study area is the mostly Mesoproterozoic Satakunta Formation (Fig. 2), which covers a northwest elongated, fault-bounded area about 15 by 100 km in size in a graben setting (e.g. Kohonen and Rämö, 2005). It is correlated to the Gävle and the Nordingrå sandstones in Sweden (Welin and Lundqvist, 1984). In the eastern Bothnian Sea area merely thin basal erosional remnants of this sandstone deposit are left, and the bedrock consists

mainly of granitic and gneissic rocks (Koistinen et al., 2001). In the northern part of the Bothnian Sea half-graben, the sandstone deposit increases in thickness, and a thickness of no less than 3 to 4 km has been suggested for the sedimentary rocks in the northern part of the Bothnian Sea (Korja et al. 2001). In the northeastern part, the formation has a subvertical, faulted contact against the Svecofennian basement (e.g. Kohonen and Rämö, 2005). The sea bottom bedrock is not known very accurately since there are no deep drillings in the study area. However, an acoustic-seismic survey provides a more locally accurate method to distinguish the Precambrian bedrock and sandstone.

Fig. 3 shows the general bathymetric information of northern Baltic Sea, and Figs. 4, 5 and 6 present more detailed bathymetric information from the Bothnian Sea study area. The discontinuous Pori–Yyteri esker,

estimated sometimes to be up to 60 m thick and 1–3 km wide (Figs. 5 and 6), continuous several tens of kilometers off the coast (Hämäläinen and Rantataro, 2001; Alvi and Kotilainen, 2002).

In the Archipelago Sea and in the northernmost Baltic Proper the crystalline rocks are exposed. They are mainly granites, gabbro, micaschist and gneisses (Koistinen et al., 2001). The bedrock of the Archipelago Sea area is fractured and there are numerous small islands and separate deposition basins.

The water depth in the study area varies up to a maximum of 50 m (Fig. 7). The steepest topographic variability in the entire study area is present in the northern Baltic Proper (Fig. 3), where the water depth typically varies between 35 and 45 m, but reaches more than 120 m in the deepest basins. Thickness of glacial and post-glacial deposits varies largely, being up to 80 m

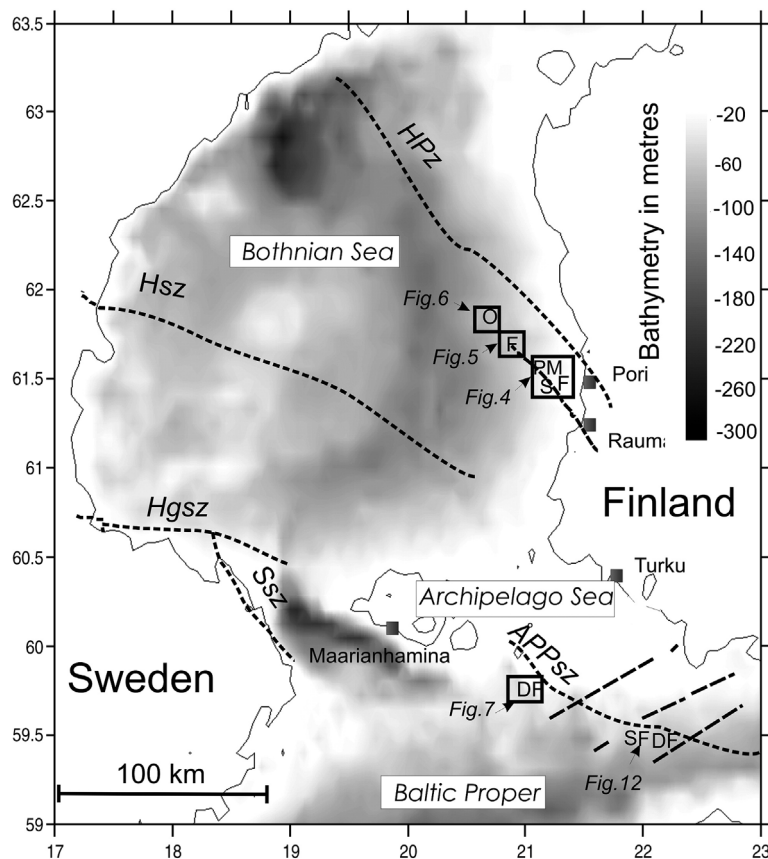


Fig. 3. General bathymetry of the northern Baltic Sea. Figure numbers refer to the maps shown in Figs. 4, 5, 6, and 7. Fig. 12 refers to echo-sounding profile shown in Fig. 12. Letters: S (=slide and slump structures), F (=faults), PM (=pockmarks and hydrodynamic blowouts), DF (=debris flows and turbidites), and O (=other structures) show the approximate locations of the observed structures in the surveyed profiles. Some major tectonic zones in the northern Baltic Sea region: Hsz = Hassela shear zone, Hgsz = Hagsta deformation zone, Ssz = Singö shear zone, ÄPPsz = Åland–Paldiski–Pskov shear zone, HPsz = Härnösand–Pori shear zone, KHHsz = Kökar–Hanko–Helsinki shear zone (Korja and Heikkinen (2005), Koistinen et al. (1996, 2001) and Flodén (1982)).

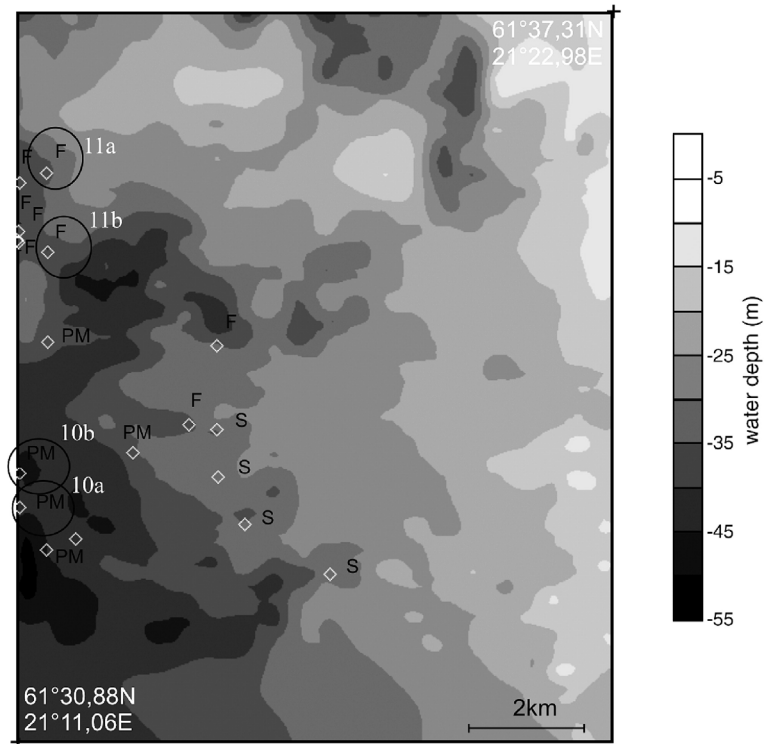


Fig. 4. Bathymetric map with locations of pockmarks, slide and slump structures, and faults (for location see Fig. 3). PM = pockmarks/hydrodynamic blowouts, F = faults and S = slides. Ellipses refer to Figs. 10a–b and 11a–b.

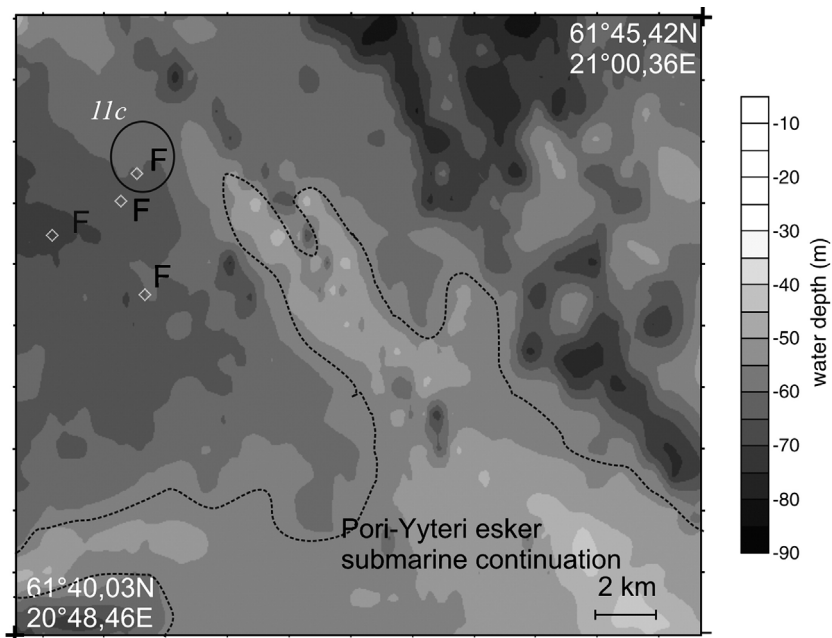


Fig. 5. Bathymetric map (for location see Fig. 3). F = faults. The dashed line indicates the assumed submarine continuation of the Pori–Yyteri esker. Numbered ellipse refers to Fig. 11c.

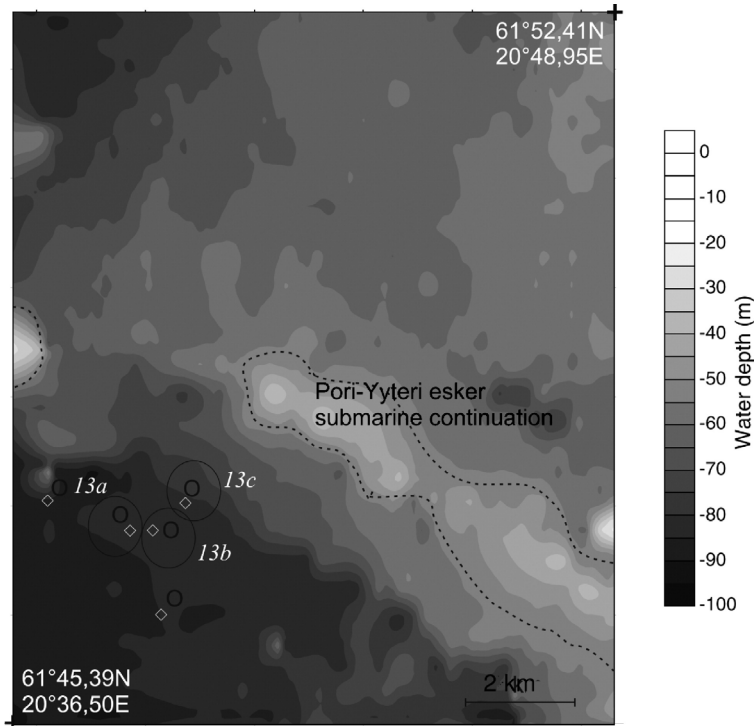


Fig. 6. Bathymetric map (for location see Fig. 3). O = unidentified structures. Numbered ellipses refer to Fig. 13a–c.

or more (Häkkinen, 1990). The most dominant features of the area are the submarine continuation of the Salpausselkä formations (Salpausselkä I and Salpausselkä II) (Fig. 2).

According to previous marine geological studies (Winterhalter, 1992; Rantataro, 2000; Kotilainen and

Hutri, 2004) the following main units can be distinguished in the northern Baltic Sea Holocene sediments. Till (unit 1) was probably deposited by the Late Weichselian ice sheet. The oldest glacioaquatic sediments (unit 2) thin upwards into distal varved sediments (unit 3) deposited during the deglaciation at the different phases

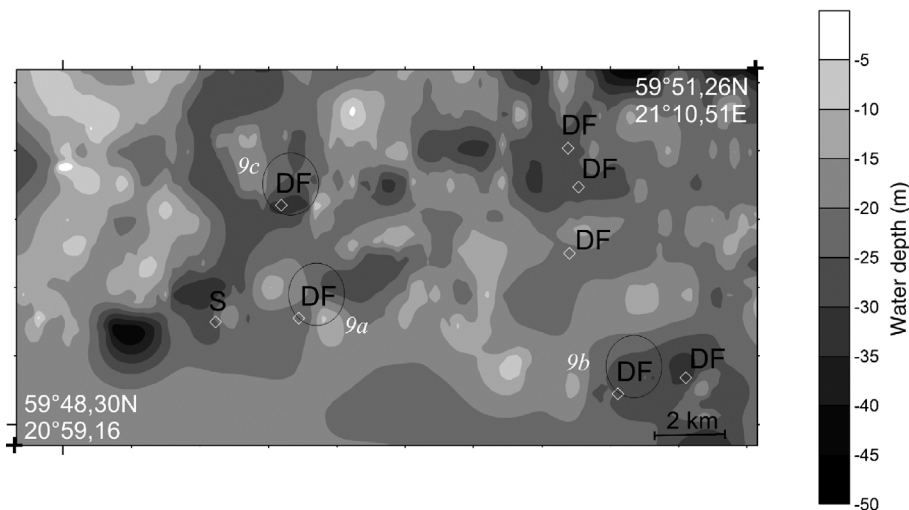


Fig. 7. Bathymetric map (for location see Fig. 3). The area contains a magnitude of shallows and small islands. DF = debris flows. Numbered ellipses refer to Fig. 9a–c.

of the Baltic Sea. Sulphide bearing clays (unit 4) are commonly associated with the “lower Ancyclus Lake” sediments (Ignatius et al., 1968). The upper parts of the unit 4, i.e. “upper Ancyclus Lake” sediments, are mainly homogenous, typically (bluish) grey clay, containing concretions of marcasite/pyrite. At the onset of the Litorina Sea stage (~8000 cal yr BP), conditions changed from freshwater to brackish (Andrén et al., 2000a,b). This change can be observed across the entire Baltic Sea area as a switch from (bluish) grey (Ancyclus) clays to organic-rich (Litorina) clay-gyttja or gyttja-clays. These sedimentary units (2 and 3) are time transgressive, i.e. older in the southern study area than in the northern study area. Litorina (unit 5) and modern sediments (unit 6) are restricted to the deep and calmer environments.

The age of the deglaciation in the study area can be estimated according to varve chronologies (Sauramo, 1929; Saarnisto and Saarinen, 2001), and the Salpausselkä I and II formations (Saarnisto and Saarinen, 2001). The final meltwater discharge of the Baltic Ice Lake to the Atlantic Ocean was a sudden event that is currently dated at $11,590 \pm 100$ cal yr BP (Saarnisto and Saarinen, 2001), being also the zero varve in the Finnish varve chronology.

3. Materials and methods

The material used in the present study consists of acoustic-seismic data including low frequency echo-sounding and single-channel reflection profiles. The survey lines are running approximately 500 m from each other, located either N–S or W–E direction (Fig. 2). Echo-sounding profiles covering 1200 km² (~2200 km of profiles) were re-examined in the present study by paying special attention to the sediment structures. All acoustic-seismic data has been recorded, and interpreted earlier, by the Marine Group of the Geological Survey of Finland (GSF). The instrumentation consisted of an MD DSS sonar system (MD DSS Multi-Mode Sonar System for Sub-Bottom Profiling, Meridata Finland Ltd). The water depth and the thickness, and internal structures, of soft sediments can be studied with the low frequency echo-sounder. Approximately 10 cm thick sediment layers could be distinguished in the present 28 kHz echo-sounder records. The digital form of the data made it possible to examine the profiles on a PC monitor. The interpretation of profiles was carried out with TOPOS mapping software (Pekkonen, 2000). The interpretations have been confirmed by several core drillings.

All bathymetric data are based on the echo-sounding profiles, but since they are located about half a kilometer from each other erroneous depth interpolations might have occurred locally in the computer calculations.

Minimum Curvature gridding method was used to reduce this problem.

In addition, a single-channel seismic reflection survey (Electro Magnetic implosion type sound source, ELMA, 400–700 Hz, depth resolution of ± 2 m) was used to examine the thicknesses and internal structures of coarse-grained sediments. The geospatial position has ± 2 m accuracy based on DGPS (Differential Global Positioning System).

In this study all sediment structures were classified according their appearance (Stow, 1994), resulting in the following five categories: 1) debris flows and turbidites, 2) slump and slide structures, 3) pockmarks and buried pockmarks, 4) faulting structures, and 5) others. The category “others” included unknown observations that could not be included with certainty in any of the other

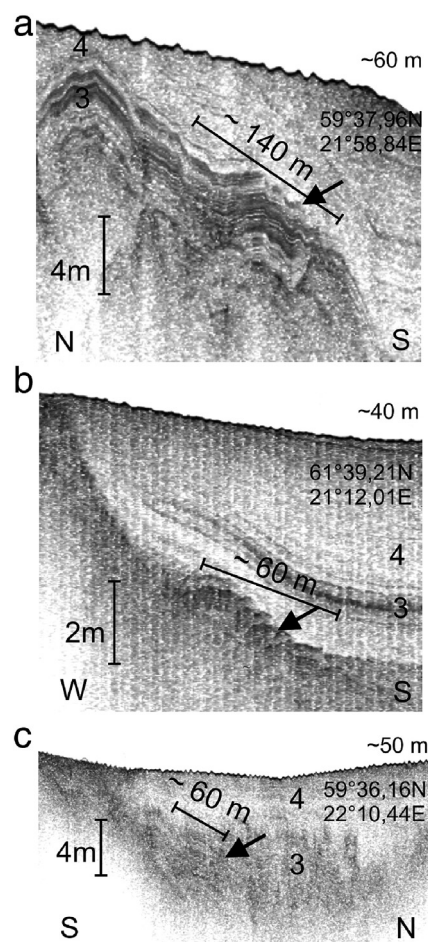


Fig. 8. a–c. Examples of slide and slump structures (indicated by black arrows). The water depth (m) and the location of the site (latitude, longitude) are given in the right corner of the figure. Numbers 3 and 4 refer to sedimentary units (3 = distal varved sediments, 4 = sulphide bearing clays) according to Rantaturo (2000).

groups. Observations that could be explained by bottom currents were omitted.

4. Observations and descriptions

4.1. Slump and slide structures

Slide and slump structures are the most common features. These are found all over the study area, located on slopes at water depths of 30–70 m. Most of these structures exist in the northern Baltic Proper, where the largest differences in topography occur (Fig. 3). Fig. 8a presents a ~140 m long shear failure (slide) along discrete shear planes with little internal deformation. In Fig. 8b the ~60 m long shear failure is accompanied by rotation along the shear planes and only weak traces of primary structures can be distinguished. The typical slide length along the profiles is 80–140 m. A slump structure ~4 m × 60 m is shown in Fig. 8c. To the right of this slump there are more structures, which may be inter-

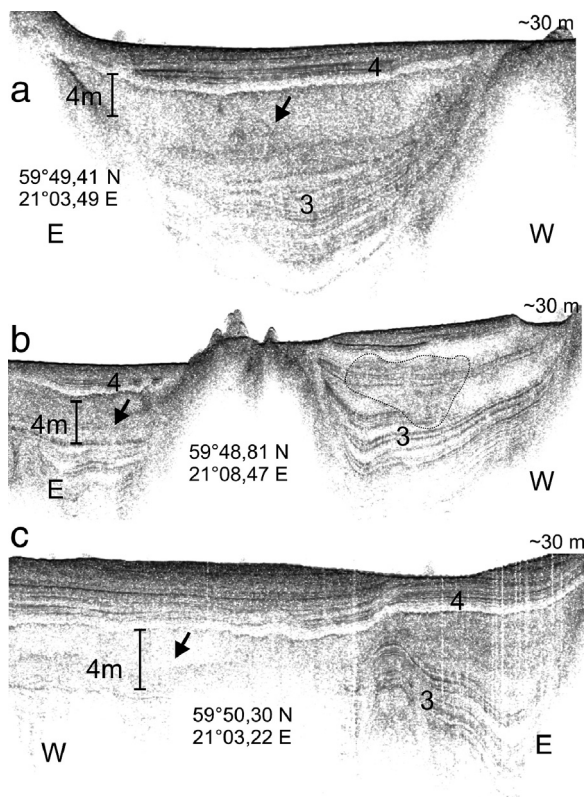


Fig. 9. a–c. Examples of the turbidite layer (indicated by black arrows). The water depth (m) and the location of the site (latitude, longitude) are given in the left corner of the figure. Numbers 3 and 4 refer to sedimentary units (3 = distal varved sediments, 4 = sulphide bearing clays) according to Rantataro (2000). A probable gas blume is marked with a dashed line.

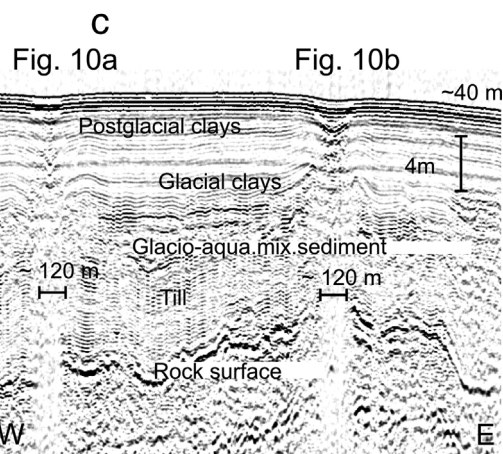
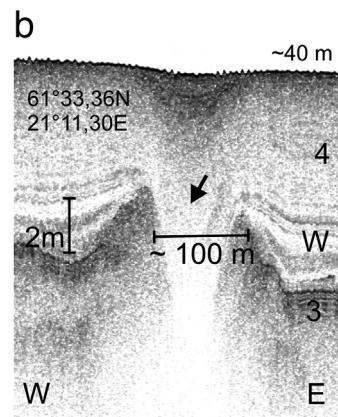
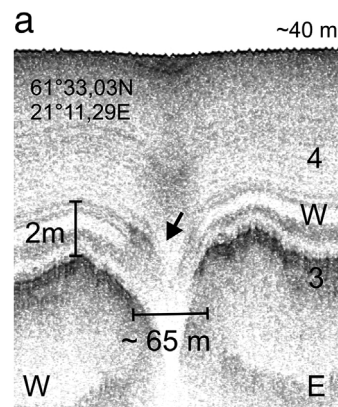


Fig. 10. a–b) The largest pockmarks/hydrodynamic blowouts (indicated by black arrows) and c) a seismic profile of them. Letters A and B refer to the Fig. 10 a, and b, respectively. The water depth (m) and the location of the site (longitude, latitude) are given in the left corner of the figure. Numbers 3 and 4, and letter W refer to sedimentary units (3 = distal varved sediments, 4 = sulphide bearing clays and W = upper Ancyclus Lake sediments) according to Rantataro (2000).

preted as gas/water structures as they have a vertical “pipe” appearance. They may be shocked sediments related to the slump. Most of these structures occur in the

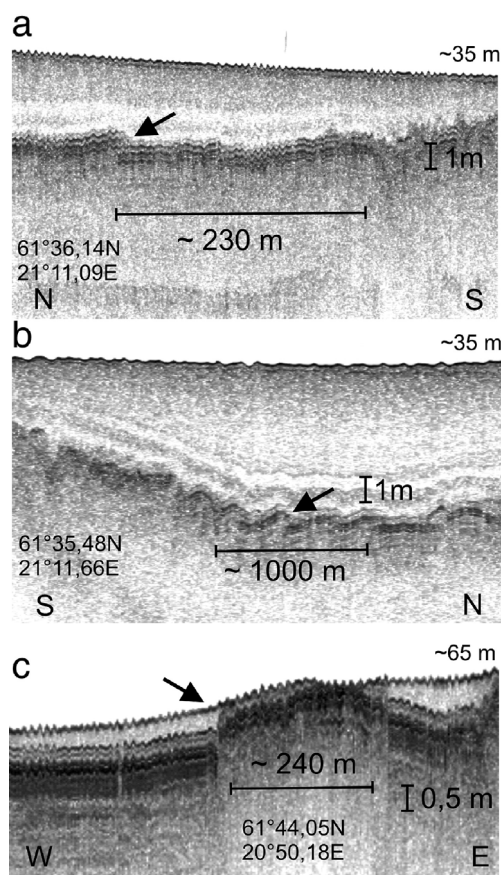


Fig. 11. a–c. Examples of faults (indicated by black arrows) in the Bothnian Sea. The water depth (m) and the location of the site (longitude, latitude) are given in the left corner of the figure. For locations see Figs. 3, 4 and 5.

uppermost part of the glacial varves and the lowermost layers of the sulphide bearing (Ancyclus) clays.

4.2. Debris flows and turbidite-type structures

In the Bothnian Sea study area (Fig. 3), debris flows only exist at a few separate sites along and below slopes in confined areas, presumably being the results of local mass flows. In the echo-sounding profile they give a chaotic signal.

A 4 to 5 m thick mixed sediment layer (Fig. 9a–c) was observed in several profiles of the northern part of the Archipelago Sea (Fig. 7), and in the basins of the northern Baltic Proper. It is typically found between the glacial varved clays and the sulphide bearing (Ancyclus) clays at the water depths greater than 30 m in the Archipelago Sea and deeper than 50 m in the northern Baltic Proper. The lower contact of the layer seems to be diffuse (Fig. 9a). Very weak internal layering can be distinguished in the

turbidite layer (Fig. 9a–c). The resolution of the profiles is not good enough to study the primary structures, but weak traces of tilted or rotated parts can sometimes be distinguished in the profile (Fig. 9c).

4.3. Pockmarks and hydrodynamic blowouts

Several pockmarks or hydrodynamic blowouts were observed in an area of $\sim 2 \times 3$ km, at a depth of 40–50 m approximately 5 km southwest from the submarine Pori–Yyteri esker (Fig. 4), in the Bothnian Sea. They are buried below ~ 5 –6 m thick clay layers below the (whitish) clay layer, usually connected to the lower Ancyclus phase. The largest structure is 120–130 m in diameter and the others range from a few meters to several tens of meters (Fig. 10a–b). In the seismic profile (Fig. 10c) the layers look like volcano craters with upward-rising rims.

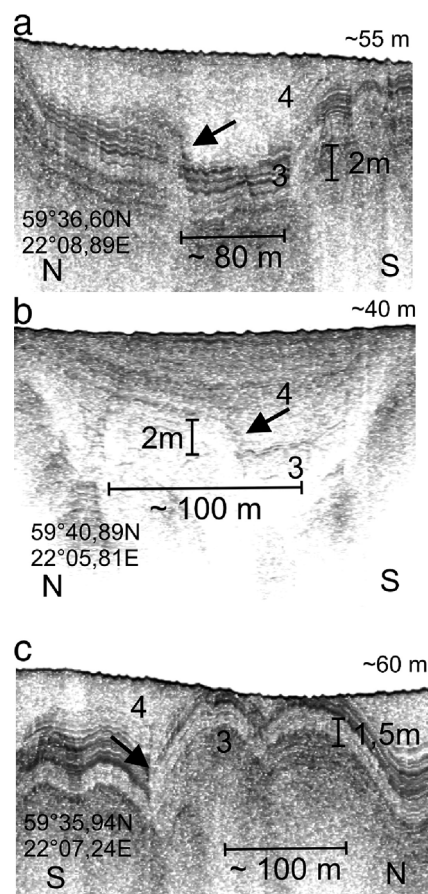


Fig. 12. a–c. Examples of faults (indicated by black arrows) in the northern Baltic Proper. The water depth (m) and the location of the site (longitude, latitude) are given in the left corner of the figure. Numbers 3 and 4 refer to sedimentary units (3 = distal varved sediments, 4 = sulphide bearing clays) according to Rantaturo (2000). For locations see Fig. 3.

According to the bedrock map and the seismic sounding, these structures are located in an area of Jotnian sediment rock trough. The bathymetric map (Fig. 4) shows that they are located on a seafloor slope. According to the fracture interpretation (Kuivamäki, 2005), the area is located between two larger fracture zones that belong to the Härnösand–Pori fault zone (Flodén, 1982; Koistinen et al., 1996, 2001) but are located about 10 km from each other.

4.4. Faulting

Faults of 0.5–1 m height (Fig. 11a–b) were found striking against 3 parallel E–W directed survey profiles in an area of $\sim 2 \times 1$ km in the Bothnian Sea study area (Fig. 4). Faulting has cut the glacial varves, but the

(Ancyclus) sediments above are not faulted. In the bathymetric map the faults are located along a quite narrow almost NW–SE trending deep (Fig. 4). According to the available seismic profiles recorded by the GSF, the underlying rock is not sandstone, contrary to the geological information given in Fig. 2, but probably igneous bedrock. Another site with a few faults (Fig. 11c) of the same size exists approximately 20 km northwest of this deep (Fig. 5). According to (available profiles recorded by the GSF) reflection seismic soundings the underlying bedrock is sandstone at this location. These faults are also located below the Ancyclus sediments.

In the northern Baltic Proper study area in the vicinity of a NW–SE trending deep, five 1.5 to 2 m high faults occur (Fig. 12a–c). At least one of the faults is located below the Ancyclus sediments, but the quality of the profiles is too weak to allow stratigraphical correlation with the others. The bedrock underneath the faults is Svecofennian crystalline rock.

4.5. Other deformation structures

In the Bothnian Sea (Fig. 6) approximately 35–40 km from the coast exist several unidentified, 2–4 m high soft sediment “heaps” (Fig. 13a–c). Since we have just single profiles across these structures they do not provide sufficient information to identify them properly. Because they could not be placed with certainty in any of the other classes they formed their own class. All heaps are located at a depth of 75–80 m, in an area of 2×4 km, and they are covered by 6 to 10 m thick clay deposits.

5. Discussion

5.1. Slides and slumps, and faults

Slide and slump structures occur along slopes throughout the study area, implying their gravitational origin during or after deposition. Slope failures (slides and slumps) may occur during deposition when the shear strength of the sediment is exceeded. They might also be induced by gas escapes from deeper layers, but there is no particular spatial relationship between them. The faults often occur together with slides and slumps, which also occur alone without faults.

The faults in the Bothnian Sea (Figs. 3, 4 and 5) discussed in this paper are located between two long fracture zones that belong to the tectonic Härnösand–Pori lineament (Flodén, 1982). They are located approximately 10 km from each other. According to fracture interpretation (Kuivamäki, 2005), no large fracture zones are located in the areas of faulting, but it must be

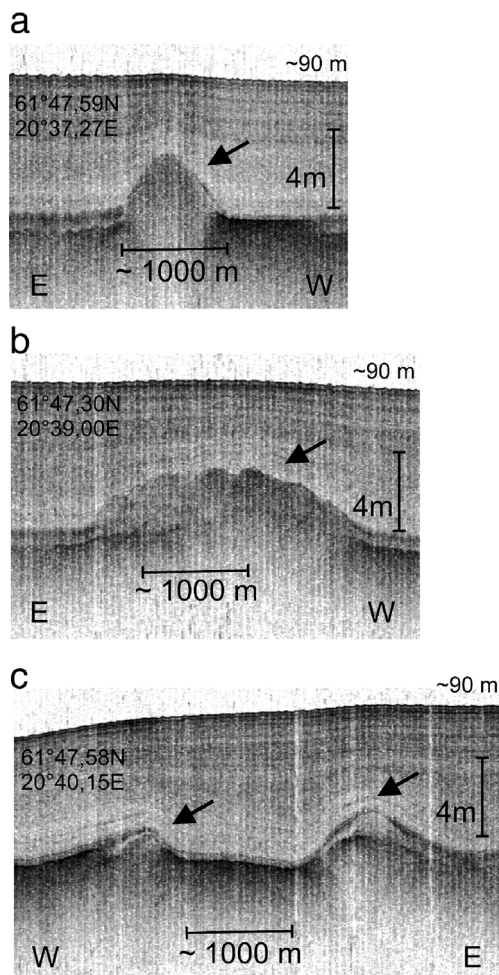


Fig. 13. a–c. Examples of the unidentified structures (indicated by black arrows). The water depth (m) and the location of the site (longitude, latitude) are given in the right corner of the figure. For locations see Figs. 3 and 6.

remembered that this interpretation is hampered by a lack of marine geological information. Since the glacial varves conform to the bottom topography, sedimentation on vertically tilted sandstones could give fault-like acoustic signals. However, although the bedrock maps (Pihlaja and Kujala, 1994, 2005; Koistinen et al., 2001) indicate that this area is comprised of mainly Mesoproterozoic sandstone (Fig. 2), the acoustic-seismic soundings do not confirm the presence of sedimentary rocks but rather crystalline bedrock.

In the bathymetric maps of the Bothnian Sea region, the faults can be associated with two depressions several hundreds of meters across with NW–SE orientation, probably developed along bedrock fracture zones (Figs. 4 and 5). One of the depressions has in addition a narrow branch in the NE–SW direction (Fig. 5). This raises the question of whether these fracture zones are affected by post-glacial movements or faulting. Post-glacial faulting in the Olkiluoto area according to different glacial scenarios has also been shown to happen by a modelling study (Hutri and Antikainen, 2002). When considering the stress field, the NW–SE oriented faulting can be connected to the ridge push from the Mid-Atlantic, whilst the NE–SW orientation coincidences with the faulting orientation due to post-glacial rebound. One possible triggering mechanism could also be gas or groundwater escapes from deeper layers.

Small (from some millimeters to 1 m) post-glacial faults with varying strike directions (almost N–S, S(S) E–N(N)W and SE–NW) have been recorded in southwestern Finland on bedrock outcrops washed by the waves of the ancient Baltic Sea (Kuivamäki et al., 1998). Ekman (1988) found that there is a spatial correlation between microseismicity and the maximum curvature of post-glacial uplift in the whole of Fennoscandia. The known earthquakes (Fig. 2) in the study area are located along NW–SE trending lineaments (Saari, 2000).

In the northern Baltic Proper, three of the faults might be associated with an old NW–SE running depression, but two faults occur several kilometers from that. This depression exists in the 150 km broad Åland–Paldiski–Pskov seismic zone. Since these faults are higher (1.5–2 m) than those observed in the Bothnian Sea subarea, they might be triggered by a more intense bedrock movement or vibration, or due to greater topographical differences. It has to be remembered that possible faults parallel to the survey profiles may be undiscovered.

Many of the slide and slump structures and all faults are found at the same stratigraphical level, i.e. the glacial varve sediments are faulted but the whitish homogenous layer associated with the lower Ancyclus (labelled with

W in the figures) is undisturbed above the faults, suggesting that faulting happened before the deposition of this layer. In the Bothnian Sea study area, sometimes also the lower Ancyclus layer was disturbed. However, it has to be taken into account that deposition of these sedimentary units happened earlier in the southern study area than in the northern study area. When compared to post-glacial faults in bedrock on land (Kuivamäki et al., 1988), the submarine faults in Holocene layers are in an advantageous position for stratigraphical correlation dating (Hutri et al., 2007).

In the Gulf of Bothnia, wave erosion in the open sea prevents the sedimentation of clay in water shallower than 50 m (Ignatius et al., 1980). Before and during the deposition of the whitish layer, the study area was submerged by more than 150 m (Eronen et al., 2001), and it is therefore unlikely that wave erosion caused the observed structures.

5.2. Debris flows and turbidites

All observations of debris flows have been in sea floor basins or on slopes. Since the observed debris flows exist at only a few separate sites in confined areas of the Bothnian Sea subarea, they presumably result from local mass flows along the slopes.

In the northern part of the Archipelago Sea, and in the northern Baltic Proper, the situation is different. The 4 to 5 m thick mixed layer (Fig. 9a, b and c), is probably a turbidite deposit. It exists on a regional scale, located under the lower Ancyclus (whitish) layer. Since the resolution of the acoustic profiles is not good enough, and no cores are available from the sites, the primary structures cannot be accurately studied. However, the diffuse lower contact of the layer and weak traces of tilted or rotated parts sometimes seen indicates an active sedimentation environment, or even distant transport (Fig. 9a). In the right part of Fig. 9b gas seems to be rising through the sediment layers and there is a gas cloud above the reflector (a “bright spot”). Below the reflector there are also indications of gas. There also might be a white “pipe” structure, which have been found also in the Stockholm Archipelago (Flodén and Söderberg 1988) with gas seepages, but the quality of the profile is not sufficient to be certain.

The northern and southern parts of the Archipelago Sea study area are separated by a shallow area (with islands) not surveyed. All debris flow observations have been north of the shallow area and only in the basins. In the western part of the northern Baltic Proper (Fig. 3) the layer is again found in the basins, adjacent to the discontinuous Salpausselkä II. Debris flows have not

been observed along the eastern side of the Salpausselkä formation, except in two basins in the southeastern part of the area. A distance of more than 15 km separates these observations, thus they might be related to separate debris flows.

A similar turbidite-type layer has also been found in Paimionlahti and Kråkskär in the northern Archipelago Sea (Virtasalo et al., 2006; Virtasalo, 2006), and the Åland Sea (Hämäläinen et al. 2005) (Fig. 2). In the Paimionlahti area the turbidite-type layer has been interpreted as the Trollskär Allomember (Virtasalo et al. 2005). The thickness of the layers found in the northern Archipelago Sea and in the Åland Sea is approximately the same as reported here (4 to 5 m), and it increases up to 15 m in the deepest basins. It appears that the thickness of the layer increases towards the contemporary ice margin, and it is therefore probable that the turbidite has come from the north/northwest. The shallow areas in the Archipelago Sea and in the northern Baltic Proper (the Salpausselkä formations) might have been controlling factors for the dispersion of the turbidite in these areas. In the Bothnian Sea study area there were no observations of a similar type of layer.

In the literature, three mechanisms are given for the initiation of turbidites (Walker, 1994): earthquakes, flooding and sediment failures in rapidly deposited delta fronts. The Weichselian ice sheet retreated from the northern Baltic Proper and the Archipelago Sea by 10,900 years BP (e.g. Strömberg, 2005). Since the turbidite layer is located between the glacial clays and the lower Ancyclus (whitish) sediments, it was deposited before or at the onset of the deposition of the Ancyclus layer (~10,700 to 9000 cal yr BP). Immediately after deglaciation the crustal rebound was at its greatest and probably caused earthquakes. In the areas where the turbidite layer was found no faults were observed. One reason for this might be that no faulting has occurred in those areas at all, or that the turbidite itself is triggered by a distant seismic event and it might have destroyed or prevented other traces of seismicity. In any case, the amount of material that has been redeposited is considerable. In eastern Canada, in the Saguenay Fjord, similar rapidly deposited layers are also connected to earthquakes over the past ~7200 years (St-Onge et al., 2004).

Other possible sources of the material could be a huge surge or the splitting off of an ice berg from the retreating Weichselian ice sheet, causing a “tsunami”. Rapid drainage of a dammed ice lake in the Bay of Bothnia could also have caused a flood related turbidite formation. The area covered with the turbidite layer is so wide that it should also have left some traces in the ancient shorelines. In Sweden, Mörner (1996) has

reported about a palaeoseismic event and an associated tsunami in 10,430 varve year BP (11,280 cal yr BP) BP causing the Närke outlet.

Changes in water level can also affect sedimentation rates. One source of the turbidite layer material could have been the rapid drainage of the Baltic Ice Lake when the water level lowered by ~25 m to the ocean level at ~11,590 cal yr BP, affecting sedimentation and redeposition in the Baltic Sea area (Veski et al., 2005). This has led to the deposition of gravel and sand layers on glacial varves near the coast of Estonia (Veski et al., 2005). However, at that time the whole study area had not yet deglaciated. The final phase of the Yoldia Sea (11 200–10,700 cal yr BP) was also strongly regressive, and the surface waters in the offshore areas e.g. of the Gulf of Finland were turbid (Heinsalu, 2001). If the water level decreased so much that the glacial varved clays were eroded, it is unlikely that a several meters thick turbidite layer could have been concurrently deposited.

Changes towards a more proximal sedimentary environment (glacier readvance) could also add to the amount of material, but there should then be glacial varves above the turbidite layer deposited in the new retreat phase. Whether the retreat of the Weichselian glacier halted for a longer time to allow delta growth at the ice margin, sediment failures of the delta fronts could also have caused turbidites, but then the deltas must have been widely distributed along the ice margin. Such a halt corresponding with the age of the turbidite layer is known to have caused the Central Finland ice-marginal formation (~11,000 years BP) within about 100 years (Sauramo, 1923; Rainio, 1996).

5.3. Pockmarks and hydrodynamic blowouts

In the northern Baltic Sea, pockmarks have so far been found, among other areas, near the island of Örö (personal information, Boris Winterhalter & Jyrki Rantataro), near Olkiluoto island (Kotilainen and Hutri, 2004), and in the Stockholm Archipelago (Flodén and Söderberg, 1988; Söderberg and Flodén, 1991, 1992). In the Stockholm Archipelago (in Sweden), the pockmarks occur along a seismically active deep crystalline rock structure. Sources of the gas are suggested to be biogenic and thermogenic (Söderberg and Flodén, 1991, 1992).

The location of pockmarks at the margins of the Bothnian Sea deeps may associate them with deeper gas seepages from the bedrock fractures. Another explanation for pockmarks could be (ground water/fluid) discharges from the Yyteri esker. However, as the esker is located ~6 km from the site, it is unlikely to have been involved

with the pockmarks. Other possible sources for gas seepages could be ancient organic material buried by till and methane clathrates formed in permafrost conditions before glaciation.

Another possible cause for the structures could have been ancient hydrodynamic blowouts triggered by the Weichselian ice sheet. In southeast Saskatchewan, Canada, the Howe Lake (diameter 295 m) is interpreted to have been formed by such a hydrodynamic blowout of water from deeper bedrock shortly after deglaciation near the ice margin in an interlobal position (Christiansen et al., 1982). The extreme overpressure of water (estimated hydraulic head 400 m) could have been induced by recharge from the melting ice sheet that had already retreated from the site. In the Bothnia Sea, the buried pockmarks are located between the glacial varved clays and the Ancyclus sediments. According to the Finnish varve chronology (Sauramo, 1929; Strömberg, 1990), the Weichselian ice sheet retreated from this area between ~10,690 and 10,890 cal yr BP. Thus, the pockmarks/hydrodynamic blowouts formed before or at the onset of the deposition of the Ancyclus sediments (~sulphide-rich clays 10,100 to 10,700 cal yr BP). In the Olkiluoto area, hyperbolic or columnar-shaped acoustic anomalies, probably caused by seeping gas bubbles, have also been identified in the water column above some of the pockmarks, suggesting on-going activity (Kotilainen and Hutri, 2004).

Another possibility is that they are “mud-volcanoes”, where material brought to the seafloor, and possibly also material in the seawater, has sedimented around the central shaft, building a hilly structure. Mud-volcanoes, with similar appearance and dimensions, but being still active, have been found in the Stockholm Archipelago (Söderberg and Flodén, 1992).

In Fig. 10a the sediment in the crater is transparent, and there are indications of some gas in the covering sediment. The sediment layering can be faintly traced through the structure and a bowl shaped depression is present in the uppermost sediment. In the sediment, inside and above the buried pockmark (Fig. 10b), the layering is not traceable through the structure. This indicates outflow mainly of groundwater and not much gas, at least not at present time. The bowl shaped sediment structures near the seafloor, above the larger structure may denote younger, recently buried, pockmarks. On the other hand, these latter structures may be diagenetic, namely chemical precipitation from the groundwater flow, slightly changing the velocity/density properties of the sediment. It is also notable that the gas in sediments lowers the sound propagation velocity and affects drastically the appearance of the sediment

structures in the profiles. A lowering of sound propagation velocity would entirely change impressions.

5.4. Others

All observed heaps (Fig. 13) are located close to each other at great depths (of 75–80 m) and covered by 6 to 10 m of clay deposits. Because the sea bottom topography underneath them is flat (Fig. 6), one explanation is that they are separate debris flows from distant slopes. Another explanation is that they are buried mud mounds formed by gas/groundwater discharges. The structure in Fig. 13a could be a mud-volcano, with the profile crossing over its side rather than its centre. Profile in Fig. 13b looks like a slump structure but it is not located at a base of some slope. The profile in Fig. 13c could image a cross-section of a central part of a mud-volcano(s) being active only for a short period because the layering above is not disturbed. On the other hand, the time has been long enough for a volcano rim to build up. There is still some gas contained in the central shaft of the mud-volcano, and also at the outer slope of the rim (left side of profile). Unfortunately, single profiles across these structures are not sufficient to identify them properly.

5.5. Future work

Post-glacial tectonic movements, high hydraulic gradients (related to glacial meltwater intrusion and discharge) and the production and discharge of deep gas are relevant to safety assessment of a spent nuclear fuel repository since they are the major threats to repository safety related to glacial scenarios (Vieno and Nordman, 1999).

Further research providing evidence to corroborate the above conclusions is still needed. More information would be gained if the profiles were closer to each other or perpendicular to the present direction of travel. Future studies extending to the ice margin of that time may reveal more information on the origin of the turbidite layer. Sediment core studies from sites with faults, turbidites and pockmarks would give more insight into their depositional history. Further investigations focusing on a porewater composition at the centre of the pockmarks could provide more information on the origin of gas seepages.

6. Conclusions

In the present study, ~2200 km of acoustic-seismic profiles were thoroughly examined to identify possible Holocene sediment structures of palaeoseismic origin.

The acoustic-seismic stratigraphy revealed new insights into the neotectonic history of the northern Baltic Sea. The observed structures included faults, slides and slumps, debris flows (turbidites), pockmarks and unidentified structures. In the study area, especially the faults showing a spatial distribution along three NW–SE or SW–NE bedrock depressions in the Bothnian Sea and in the northern Baltic Proper are suggested to be triggered by palaeoseismic event(s). The slide and slump structures that occur in the same stratigraphical position in the vicinity of the faults are probably of the same origin. The 4–6 m thick turbidite-like layer found in the same stratigraphical position in the Archipelago Sea and in the northern Baltic Proper might also be caused by palaeoseismicity. Other possible sources such as a massive surge from the retreating Weichselian ice sheet, a rapid drainage of a dammed ice lake or Yoldia regression could also have caused the turbidite(s). The pockmarks might be related to bedrock fractures that were seismically reactivated after deglaciation, or they might have been formed due to high hydraulic gradients connected to the meltwaters of the Weichselian ice sheet.

All the observations that could be stratigraphically dated were formed between the glacial distal varves and the “lower Ancylus Lake” sediments (from ~10,100 to ~10,700 cal yr BP), getting younger to the northern parts of the study area. All evidence of post-glacial faults indicates that seismic reactivation has been located along the old fracture zones. Since no younger or repeated traces of possible seismic events were found, it corroborates the suggestion that seismic activity in the area happened within a short time during and after deglaciation.

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