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Southwestern Finland was covered by the Weichselian ice sheet and experienced rapid glacio-isostatic rebound after early Holocene deglaciation. The present mean overall apparent uplift rate is of the order of 4–5 mm/yr, but immediately after deglaciation the rate of crustal rebound was several times higher. Concurrently with land uplift, relative sea level in the Baltic basin during the past more than 8000 years was also strongly affected by the eustatic changes in sea level. There is ample evidence from earlier studies that during the early Litorina Sea stage on the southeastern coast of Finland around 7000 yr BP (7800 cal. yr BP), the rise in sea level exceeded the rate of land uplift, resulting in a short-lived transgression. Because of a higher rate of uplift, the transgression was even more short-lived or of negligible magnitude in the southwestern part of coastal Finland, but even in this latter case a slowing down in the rate of regression can still be detected. We used evidence from isolation basins to obtain a set of 71 ^{14}C dates, and over 30 new sea-level index points. The age-elevation data, obtained from lakes in two different areas and located between *c.* 64 m and 1.5 m above present sea level, display a high degree of internal consistency. This suggests that the dates are reliable, even though most of them were based on bulk sediment samples. The two relative sea-level curves confirm the established model of relatively gradually decreasing rates of relative sea-level lowering since *c.* 6100 yr BP (7000 cal. yr BP) and clearly indicate that the more northerly of the two study areas experienced the higher rate of glacio-isostatic recovery. In the southerly study area, changes in diatom assemblages and lithostratigraphy suggest that during the early Litorina Sea stage (8300–7600 cal. yr BP) eustatic sea-level rise exceeded land uplift for hundreds of years. Evidence for this transgression was discovered in a lake with a basin threshold at an elevation of 41 m above sea level, which is markedly higher than any previously known site with evidence for the Litorina transgression in Finland. We also discuss evidence for subsequent short-term fluctuations superimposed on the main trends of relative sea-level changes.

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Glacio-isostatic rebound of the earth's crust has been the determining factor in the late and post-glacial development of the Baltic basin. The uplift was extremely rapid during and immediately after Late Weichselian deglaciation and it continues today at rates up to 9 mm/year in the central areas affected by the Weichselian ice sheet (Fig. 1; Vermeer *et al.* 1988; Ekman 1989, 1996). Owing to uplift and consequent tilting of the land, several water connections between the Baltic Basin and the open ocean existed and closed at various locations and times during the late Weichselian and Holocene. Successive episodes of fresh and saline water characterize the eventful history of this inland sea (Fig. 2).

The most common type of evidence traditionally used to reconstruct the various shore levels has been geomorphological shoreline features. Tens of shoreline displacement curves have been drawn and published for different areas in Finland (Eronen *et al.* 1993, 1995) and a diversity of shoreline diagrams can thus be used to reconstruct the raised shorelines. Most of the earlier curves suffer from poor resolution or imprecise geochronological information. Since the 1970s, the chronology for the curves has been mostly based on radiocarbon dating (formerly it was based on pollen analysis), but most curves are based on insufficient numbers of radiocarbon dates.

Published radiocarbon-dated relative sea-level index points, disseminated through a large number of publications, have been reviewed by Eronen *et al.* (1995). Those data consistently indicated very rapid land uplift during the early Holocene times followed by deceleration of the rate of glacio-isostatic recovery (Glückert 1976; Eronen 1983, 1990; Matiskainen 1989; Eronen & Ristaniemi 1992; Donner 1995; Ristaniemi *et al.* 1997). The geographical distribution of all 259 ^{14}C dates,

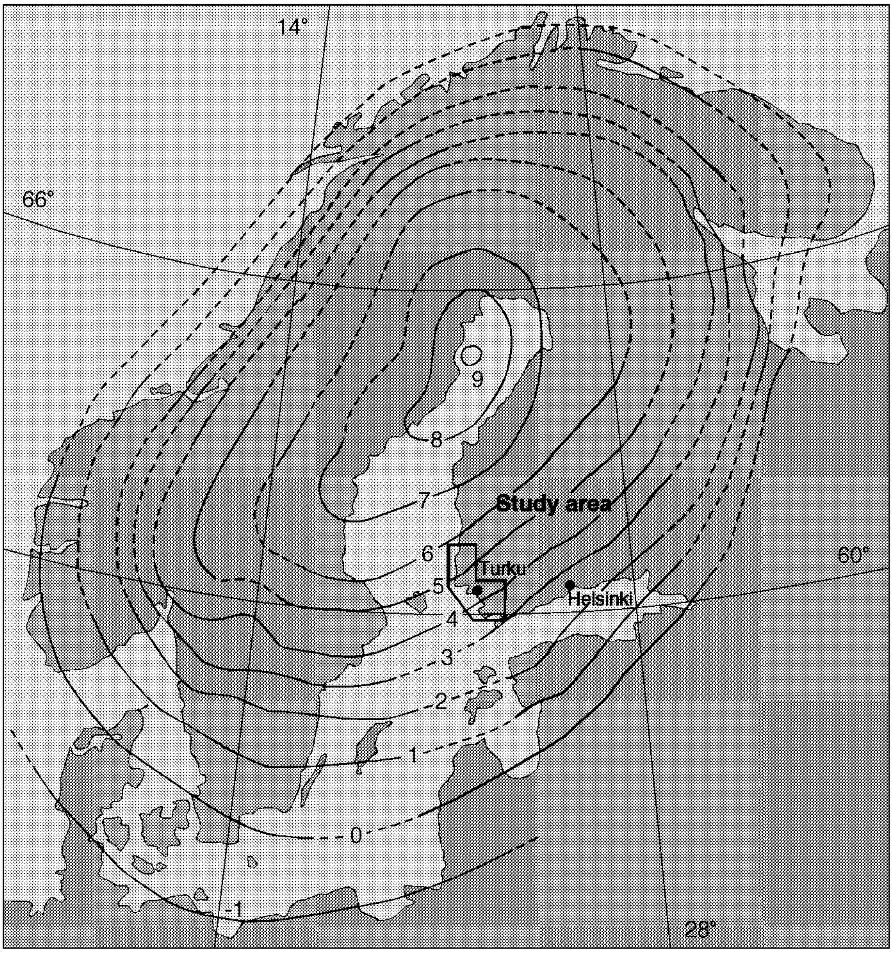

Fig. 1. Isobases (mm/yr) of present land uplift in Fennoscandia constructed on the basis of precise (instrumental) levelling, gravity and sea-level data (redrawn from Ekman & Mäkinen 1996).

obtained from 198 uplifted lake basins, is shown in Fig. 3. The largest concentration of dated samples was recovered from sites in the southern and southwestern parts of Finland, owing to the fact that here one can study palaeo-shore levels as old as the late glacial period. Moreover, an understanding of sea-level history is especially critical for this region in view of its rich legacy of archaeological findings. Note that Fig. 3 shows only the sites with radiometrically dated horizons. The overall number of sites where past shore levels were studied is many times larger. Nowadays the most precise information on Holocene shoreline displacement is obtained from small uplifted lake and bog basins, where the emergence of the basin from the sea or large lake can be detected in the sediment by changes in diatom assemblages, these 'index points' being datable by the radiocarbon method. This method is especially useful in the Baltic Sea area, where the tidal range is negligible. In Finland alone there are more than 187000 lakes providing potential for this kind of study (Atlas of Finland 1986).

In spite of the large concentration of radiocarbon-dated sea-level index points in southwestern Finland, their density varies markedly and there is considerable room for improvement of both the geographical distribution and the spatial and temporal resolution of the database. We therefore collected new data from the Tammisaari–Perniö area in 1992 and 1993, and from the Olkiluoto–Pyhäjärvi area in 1994 and 1995 (Fig. 3). The two objectives of this study were: (1) to document in considerable detail the course of relative sea-level lowering in the two study areas, and (2) to analyse and compare these records in order to recognize significant glacio-isostatic and eustatic trends during the past 8000 years.

Mastogloia phase and Litorina Sea stage

The early history of the Baltic Sea, from the late Weichselian Baltic Ice Lake, through the Yoldia Sea stage to the giant Ancylus Lake, has been discussed in numerous publications (e.g. Eronen 1983, 1990; Svensson 1989; Donner 1995; Björck 1995; Påsse 1996).

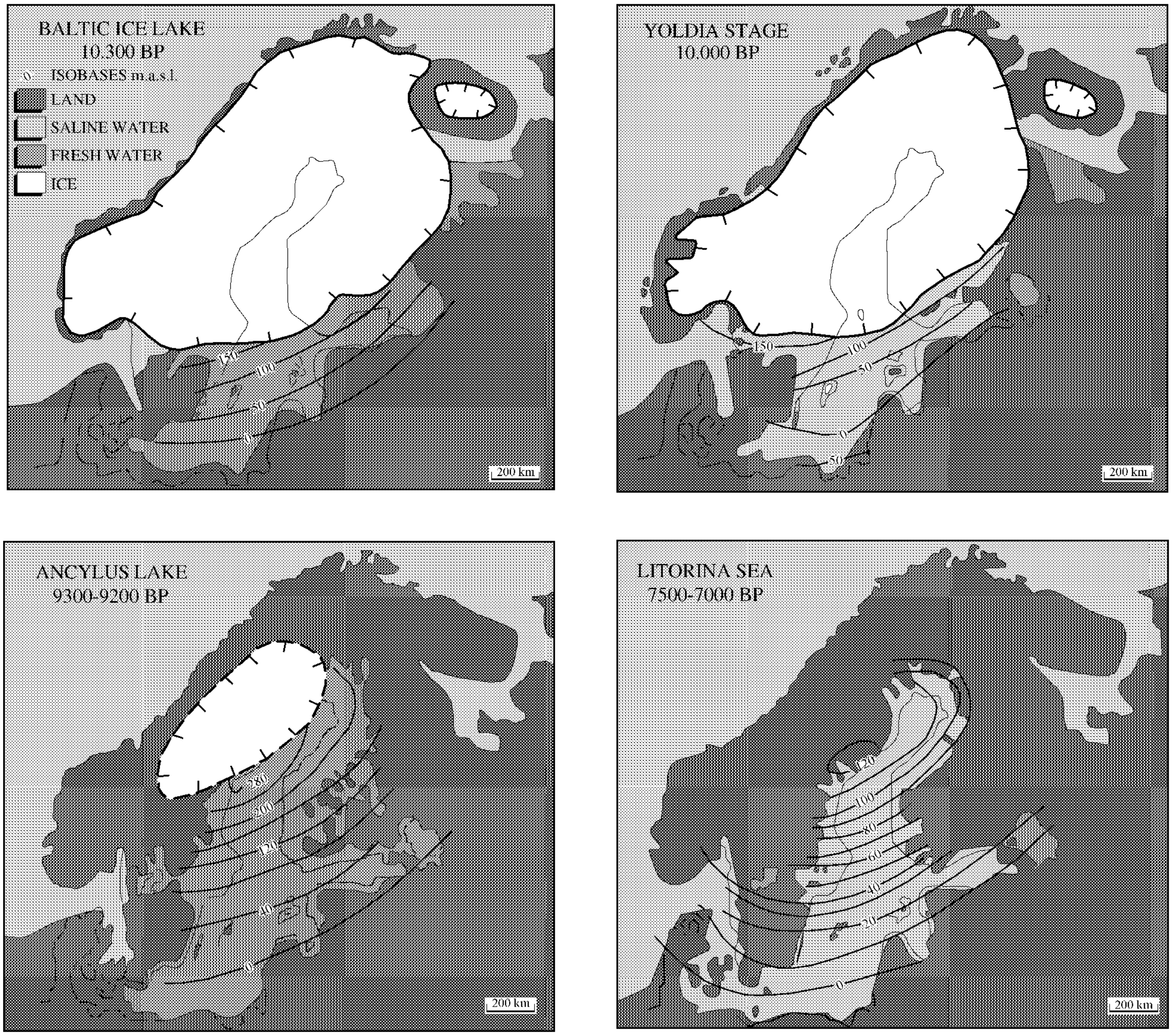

Fig. 2. The four stages in the late and post-Weichselian history of the Baltic Sea.

Because our new data do not extend back to those early stages of the development of the Baltic Sea, they are not discussed in detail here.

A new phase in the history of the Baltic began when the Ancylus Lake stage came to an end after the rapid drainage to the ocean in the southwestern part of the Baltic basin (Winn *et al.* 1986; Eronen *et al.* 1990; Björck 1995). By this time the altitude of the surface of the Baltic approximated that of the open ocean and the flow of saline water gave rise to a transitional, slightly saline phase, the Mastogloia phase, before the more brackish Litorina Sea stage commenced. Because the water level in the Baltic basin has since that time been very close to the level of the open ocean, the world sea-level changes must have affected the shoreline displacement of the Baltic during the past *c.* 8500 ^{14}C years.

'Mastogloia' are diatoms that commonly live in slightly saline waters. Species of *Mastogloia* characterize the intermediate phase, when an influx of marine water began to alter the large-lake ecosystem of the Baltic basin. The term "Litorina Sea" dates back to the 19th century, when shells of species of the brackish water gastropod *Littorina littorea* were discovered in raised beaches in Sweden (Lindström 1886).

The saline ocean water spread to the Baltic basin from the southwest, flowing first into the deeper southern parts. The conversion to brackish conditions of the entire Baltic basin probably took several hundred years and thus the change from fresh-water to brackish

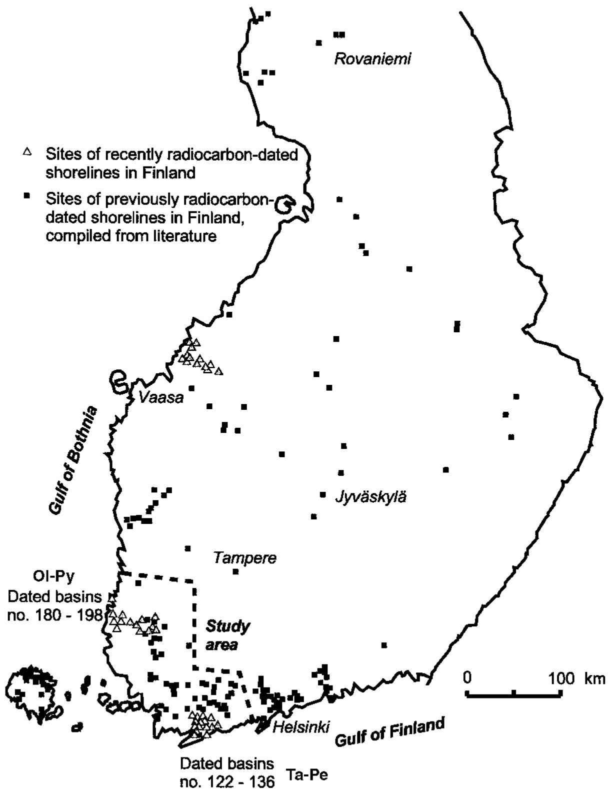

Fig. 3. Sites of radiocarbon-dated shorelines in Finland, encompassing 198 basins and 259 dates (Eronen *et al.* 1993, 1995).

conditions in the basin was time-transgressive. The weak salinity of the Mastogloia phase can be detected in the southwest Baltic basin by 8200 BP (Björck 1995). The earliest record on the Finnish coasts is *c.* 8000 BP (Hyvärinen 1984). Most of the Baltic was distinctly brackish by 7500 BP, which marks the onset of the Litorina Sea stage. In the Gulf of Bothnia area, in the northernmost part of the Baltic, this change was delayed by *c.* 500 years (Eronen 1974). A delay of the same duration has been found in studies of sea-level history in central Sweden (Hedenström & Risberg 1999).

The change to brackish conditions in the Baltic also brought an increase in nutrient concentrations and so the Litorina Sea became more eutrophic than during earlier stages. Thus the onset of the Litorina Sea is an important marker horizon, which can be observed in the lithological record as a change from grey Ancylus (including Mastogloia) clays to organic-rich Litorina clay-gyttjas or gyttja-clays (Winterhalter *et al.* 1981). A similar change is indicated in the diatom stratigraphy by the replacement of assemblages dominated by freshwater species by markedly different brackish-water assemblage. The upper limit of the Litorina Sea stage can therefore be easily detected and dated by determining the upper limit of brackish-water sediments in the raised shore levels (Eronen 1974, 1990).

At the start of the Litorina Sea stage, transgression resulted from the rising global sea level. In fact, the eustatic rise caused by the melting of the world's ice sheets was already underway during the Mastogloia phase. This submerged a basin threshold in Denmark, and thus the rising ocean should also have caused a "Mastogloia transgression" in large parts of the Baltic. That episode is difficult to detect, however, because any deposits associated with that event are immediately overlain by the brackish-water deposits of the subsequent, more emphatic Litorina transgression. The rising water flooded into those areas where the rate of uplift was slower than the rate of sea-level rise. On the coasts of Finland, a marked decrease in the rate of uplift can be observed between 8500 and 8000 BP, allowing the sea to rise relative to land in southern and southeastern Finland during *c.* 1000 years after the beginning of the Litorina Sea stage (Eronen 1983, 1990). This transgression petered out when the eustatic rise slowed down to a halt by 6000–5000 BP (Pirazzoli 1991). During the past 6000 years, relative sea level has fallen progressively due to land uplift, while, at the same time, salinity in the Baltic has decreased as a result of narrowing of the connection to the ocean (Glückert 1976; Eronen 1974, 1983, 1990). It is commonly agreed that the Litorina Sea stage ended, and the less saline Limnaea substage began, at *c.* 4000 BP (Donner 1995), even though no sharp boundary can be distinguished.

Study areas

Sediment cores were collected from lakes in two areas in southwestern Finland. In the first phase of this study (1992–1993), cores were collected from the Tammisaari–Perniö (Ta-Pe) area (Fig. 3). That region proved highly suitable for studies on relative sea-level changes due to the abundance of small lakes and peat bogs in depressions between numerous rocky hills lying at varying elevations a.s.l. Later on, data were needed for a study of relative sea-level lowering in the vicinity of the nuclear power plant at Olkiluoto. For this reason, in 1994–95 sediment cores were collected in the Olkiluoto–Pyhäjärvi (Ol–Py) area and thus the investigations were extended into two different areas (Eronen *et al.* 1993, 1995).

The Ta–Pe area covers about 90 × 60 km and the Ol–Py area about 70 × 60 km. The bedrock in both regions, as elsewhere in southwestern Finland, consists mainly of early Proterozoic metamorphic and igneous rocks (age > 1.8 Ga). In the Ol–Py area, younger Proterozoic rocks, Rapakivi granite (age 1.6 Ga), Jotnian sandstone (1.35 Ga) and diabase dykes (1.31 Ga) also occur (Laitakari 1925; Härme 1958; Hämäläinen 1987). Because of the low relief in the sandstone and Rapakivi areas (Tikkanen 1981), the Ol–Py region has far fewer small lakes than the Ta–Pe area, and is less but nevertheless still well suited for detailed studies of the history of shoreline displacements.

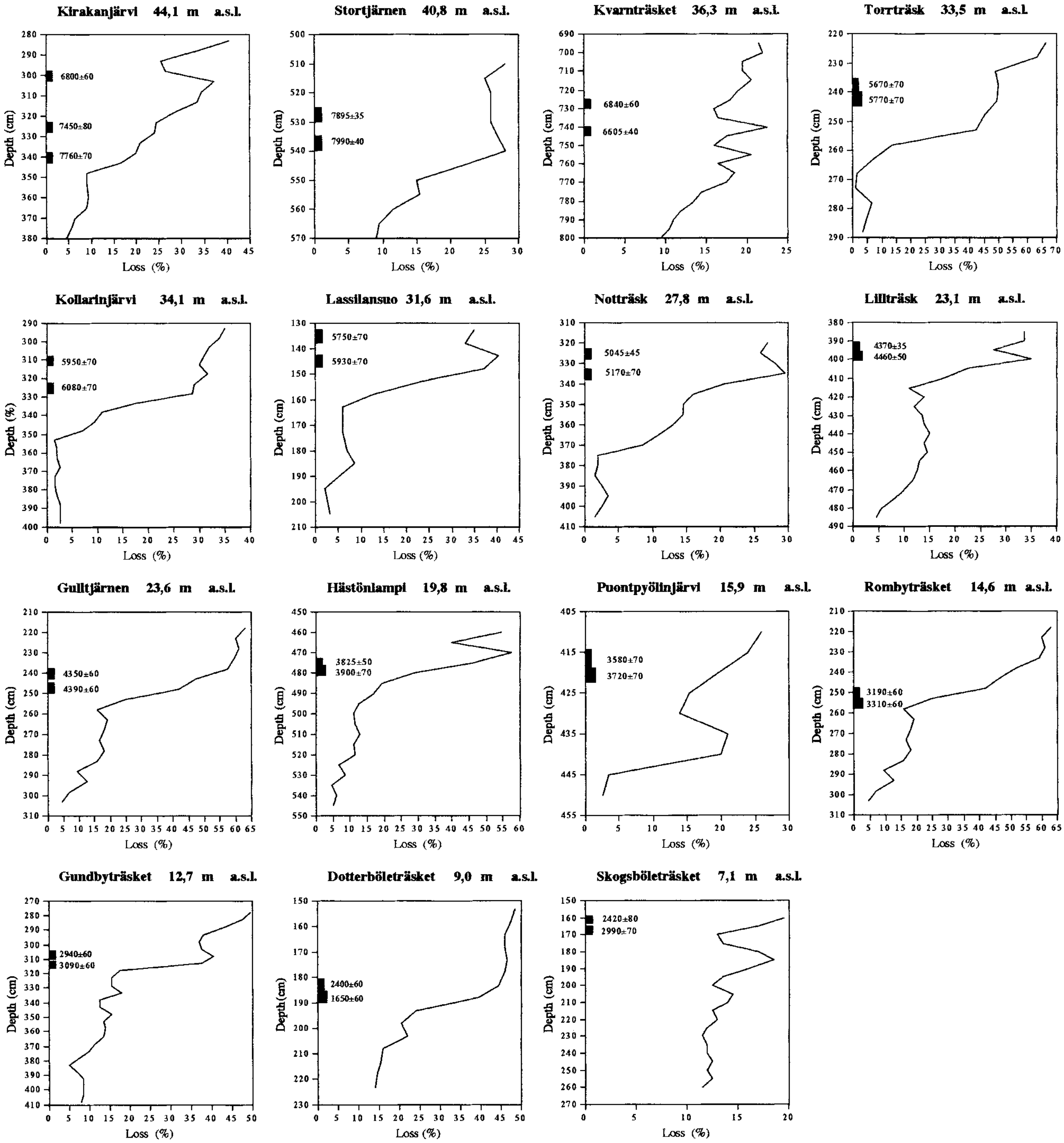

Fig. 4. Loss-on-ignition curves from 15 dated basins in the Tammisaari–Perniö area. A bulk sediment dating is made for the beginning and end of isolation in each basin. For Lake Kirakanjärvi one additional date was obtained of a sediment layer indicating a transgressive phase (cf. text).

Both areas were submerged by deep water following regional deglaciation in early Holocene to emerge later due to isostatic rebound. The measured rates of present-day land uplift range from 4 to 5 mm/year in the northwest (Ol–Py area) to 3 to 4 mm/year in the southeast (Ta–Pe area). The values include a correction for an assumed steady 0.8 mm/year rise in global sea level (Kääriäinen 1966; Suutarinen 1983). The figures for *apparent* uplift (relative to sea-level movement) are correspondingly lower (Ekman 1989, 1996).

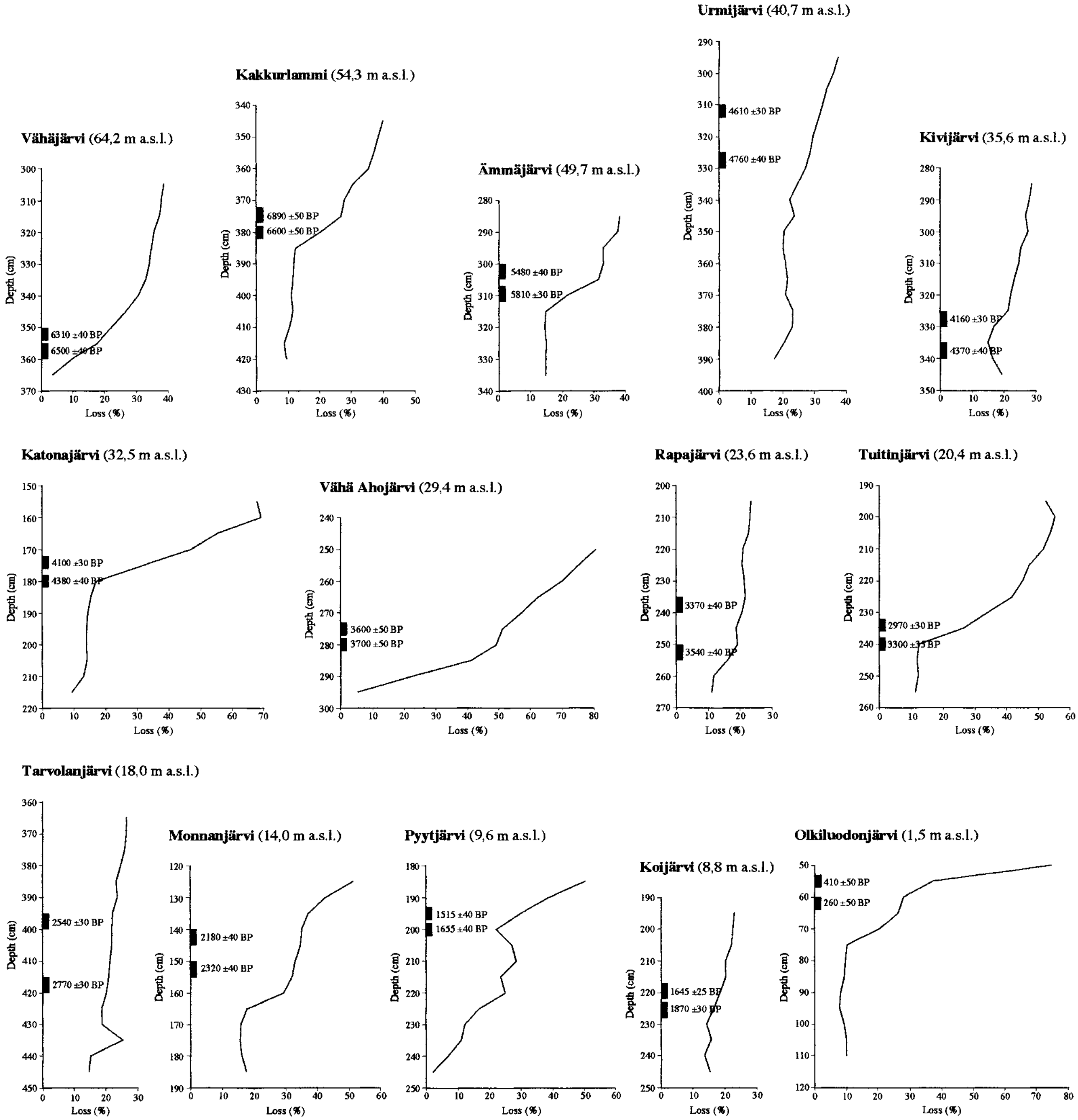

Fig. 5. Loss-on-ignition curves from 14 dated basins in the Olkiluoto–Pyhäjärvi area (cf. Fig. 4).

Because of differential land uplift, ancient shorelines in the area are tilted in a NW–SE direction. The highest Litorina shoreline descends from 70 m to 58 m a.s.l. in the Ol–Py area and from 48 m to 38 m in the Ta–Pe area (cf. Fig. 2). The extrapolated altitudes of the highest postglacial strandlines in these areas are about 145 m and 130 m a.s.l., respectively. In the Ta–Pe region the highest shore level was formed during the Baltic Ice Lake, whereas that in the Ol–Py area was formed in the Yoldia stage of the Baltic (Eronen 1974, 1983, 1990; Eronen *et al.* 1993, 1995).

Field work

Sites suitable for coring were located using base maps at a scale of 1:20 000. Most lakes were cored at their centre on winter ice, but some were cored during the summer, from the blanket of peat extending from the mineral ground at the shore towards the open water. Although in the latter cases the cores could not be obtained from the central part of the basins, nevertheless they do appear to contain evidence of an isolation horizon. In one case (Dotterböleträsket), additional samples were taken in the subsequent winter, to verify that the isolation horizon had been correctly determined.

A Russian peat sampler (length 600 mm, diameter 77 mm) and a piston corer (length 700 mm, diameter 80 mm) were used in the collection of sediment cores. The lithostratigraphy of the lake sediments was examined preliminarily and the isolation horizon was noted by visual inspection. Usually it is readily detected as a change upwards from greyish sediments deposited during submergence to greenish or brownish lake gyttja deposited after emergence. One-metre long core segments extending below and above the isolation contact were collected from each lake for detailed laboratory analysis. Water depths in all sampled lakes did not exceed a few metres.

Although the base maps give the approximate altitude of each lake, the elevations of the outlet thresholds were determined by instrumental levelling with an accuracy of ±5–10 cm. In several cases, the original threshold had been artifically lowered, sometimes even by 2 m, by ditch excavations in the outlet. Ditch construction for the purpose of increasing the area of productive meadows was very common during the 19th century and in the early part of the 20th (Anttila 1967). It was possible to estimate the magnitude of the consequent drops in lake levels by field observations (with an estimated accuracy ±20 cm) and its effect was taken into account in defining the elevations of the thresholds of the basins used in the present study.

Laboratory analyses

The sediment cores were studied at the Department of Quaternary Geology, University of Turku. Loss-on-ignition was determined by drying the samples for 6 h at 105°C and subsequent combustion at 550°C. Samples for diatom identifications (1 cm^3) were taken from the cores at intervals of 5 cm. Hydrogen peroxide (30%) was used in order to bleach and destroy the organic matter. The coarse mineral grains were removed by repeated decanting and diluting the diatom fraction suspended in water. Adequate amounts of condensed suspension were transferred to cover slips and permanent mounts made in Canada balsam.

The organic content of the sediment was measured by loss-on-ignition as an indicator of the level representing lake basin isolation (Figs 4 and 5). As a rule, the sediment deposited in small lakes is organic-rich gyttja and thus the onset of its deposition in a core indicates that the basin must have emerged from the sea. More conclusive evidence for a change from marine or brackish environments to a fresh-water lake was provided by the diatom analysis. For the present study, only preliminary diatom assemblage data were picked up to define the isolation contacts and to establish the appropriate segments of each core suitable for radiocarbon dating. The dominant species indicating brackish or fresh-water conditions were identified. Such preliminary studies led to the rejection of some cores, because of evidence of hiatuses or perturbations in the sediments. This study is based on the remaining 33 cores which survived this initial screening process.

From each core retained for further study, blocks of sediment of 4–6 cm vertical thickness were cut from the uppermost brackish-water sediment unit and from the lowermost portion of fresh-water gyttja for radiometric dating. Consequently, two limiting ^{14}C dates are available for each isolation horizon. In five of the lake basins, one additional limiting date was obtained (Table 1). Most of the dates are based on bulk samples. Four AMS radiocarbon dates were obtained from seeds or plant fragments extracted from the sediment. For three of these, the dated material was obtained from the horizon immediately overlying that from which the bulk date had been obtained, but in one case (Lake Dotterböleträsket) a new core was available and the sample was taken precisely from the isolation contact as defined on lithological and biostratigraphical grounds. The use of bulk sediment dates was unavoidable, because the sediment at the isolation contact was usually fine detritus gyttja in which it was difficult to find any coarse plant remains. Altogether, 67 conventional and 4 AMS radiocarbon dates were used in the construction of the two shoreline curves (Figs 6 and 7).

All radiocarbon dates were measured at the Centre for Isotope Research, University of Groningen, The Netherlands.

Defining the isolations

The isolation of a basin from the sea is a gradual process which may require tens to hundreds of years depending on the rate of uplift and local shore facies. Even in areas with negligible tidal range, the local sea level may vary considerably due to changing meteorological conditions. For instance, the sea level in the gulfs of the Baltic Sea today can vary locally by 2.8 m between a low level under high atmospheric pressure conditions, and a high level during subsequent cyclonic low pressure (Atlas of Finland 1986). Such short-term fluctuations of sea level can enable saline water to enter recently isolated lakes which today lie only slightly above mean sea level. On the Finnish coast, there is even a special term used to refer to those lakes, which are still brackish but slightly above sea level, namely "glo", while entirely fresh-water basins are called "flada" (Munsterhjelm 1987). It is highly likely, therefore, that the basins isolated in the distant past must have gone through equivalent phases

Table 1. Radiocarbon-dated isolation horizons in lake basins in southwestern Finland in the Tammisaari–Perniö (basins no. 122–136) and Olkiluoto–Pyhäjärvi areas (basin nos. 180–193 and 195–198). The ^{13}C values of the materials from which the dates were obtained vary between −19 and −32‰, which is commonly seen in gyttja deposits.

	Basin, location	Coordinates	Threshold m. a.s.l.	Conv. yr BP	Cal. yr BP	Lab. no.	d13 C (‰)	Explanations	References
64.	Bastukärr, Pohja	60°06′N 23°37′E	38	7070 ± 90	8010–7770	Su-1536	-	Beginning of Litorina transgr.	Ristaniemi & Glückert 1988
122.	Kirakanjärvi, Perniö	60°12′N 22°59′E	44.5	7760 ± 70	8590–8430	GrN-19635	−19.16	Isolation begins	Eronen *et al.* 1993
	Kirakanjärvi, Perniö			7450 ± 80	8290–8130	GrN-19634	−19.04	Litorina transgr. begins	Eronen *et al.* 1993
	Kirakanjärvi, Perniö			6800 ± 60	7666–7562	GrN-19633	−23.14	Litorina transgr. ends, isol.	Eronen *et al.* 1993
123.	Stortjärnen, Pohja	60°04′N 23°29′E	39.9	7990 ± 40	8890–8750	GrN-19637	−23.40	Beginning of isolation	Eronen *et al.* 1993
	Stortjärnen, Pohja			7895 ± 35	8730–8630	GrN-19636	−23.71	End of isolation, Litorina	Eronen *et al.* 1993
AMS	Stortjärnen, Pohja			7630 ± 60	8450–8310	GrA-2542	−26.78	Isolation Litorina	Eronen *et al.*; this paper
124.	Kvarnträsket, Tenhola	60°02′N 23°09′E	38.5	6605 ± 40	7482–7410	GrN-19940	−25.05	Beginning of isolation	Eronen *et al.* 1993
	Kvarnträsket, Tenhola			6840 ± 60	7710–7590	GrN-19939	−25.51	End of isolation, Litorina	Eronen *et al.* 1993
AMS	Kvarnträsket, Tenhola			6230 ± 50	7150–7030	GrA-3015	−25.39	Isolation Litorina	Eronen *et al.*; this paper
125.	Torrträsk, Tenhola	60°08′N 23°13′E	35.3	5770 ± 70	6670–6510	GrN-19652	−28.31	Beginning of isolation	Eronen *et al.* 1993
	Torrträsk, Tenhola			5670 ± 70	6550–6390	GrN-19651	−28.04	End of isolation, Litorina	Eronen *et al.* 1993
126.	Kollarinjärvi, Perniö	60°12′N 23°20′E	35.0	6080 ± 70	7010–6850	GrN-19639	−27.61	Beginning of isolation	Eronen *et al.* 1993
	Kollarinjärvi, Perniö			5950 ± 70	6870–6690	GrN-19638	−30.40	End of isolation, Litorina	Eronen *et al.* 1993
127.	Lassilansuo, Pohja	60°07′N 23°27′E	33.0	5930 ± 70	6850–6670	GrN-19641	−25.87	Beginning of isolation	Eronen *et al.* 1993
	Lassilansuo, Pohja			5750 ± 70	6630–6470	GrN-19640	−26.69	End of isolation, Litorina	Eronen *et al.* 1993
128.	Notträsk, Tenhola	60°08′N 23°15′E	29.5	5170 ± 70	6010–5850	GrN-19643	−27.52	Beginning of isolation	Eronen *et al.* 1993
	Notträsk, Tenhola			5045 ± 45	5850–5750	GrN-19642	−30.54	End of isolation, Litorina	Eronen *et al.* 1993
129.	Lillträsk, Tenhola	60°05′N 23°15′E	24.7	4460 ± 50	5150–4990	GrN-19942	−30.32	Beginning of isolation	Eronen *et al.* 1993
	Lillträsk, Tenhola			4370 ± 35	5010–4910	GrN-19941	−29.60	End of isolation, Litorina	Eronen *et al.* 1993
130.	Gulltjärnen, Tammisaari	59°59′N 23°22′E	24.2	4390 ± 60	5070–4890	GrN-19646	−26.00	Beginning of isolation	Eronen *et al.* 1993
	Gulltjärnen, Tammisaari			4350 ± 60	5010–4830	GrN-19645	−29.35	End of isolation, Litorina	Eronen *et al.* 1993
131.	Hästönlampi, Perniö	60°08′N 23°04′E	20.3	3900 ± 70	4410–4210	GrN-19944	−28.78	Beginning of isolation	Eronen *et al.* 1993
	Hästönlampi, Perniö			3825 ± 50	4290–4130	GrN-19943	−32.00	End of isolation, Limnaea	Eronen *et al.* 1993
132.	Puontpyölinjärvi, Tenhola	60°05′N 23°10′E	18.2	3720 ± 70	4170–3970	GrN-17287	−27.89	Beginning of isolation	Eronen *et al.* 1993
	Puontpyölinjärvi, Tenhola			3580 ± 70	3970–3770	GrN-17286	−29.22	End of isolation, Limnaea	Eronen *et al.* 1993
133.	Rombyträsket, Tenhola	60°01′N 23°15′E	15.6	3310 ± 60	3610–3450	GrN-19650	−22.80	Beginning of isolation	Eronen *et al.* 1993
	Rombyträsket, Tenhola			3190 ± 60	3470–3310	GrN-19649	−28.66	End of isolation, Limnaea	Eronen *et al.* 1993
134.	Gundbyträsket, Tenhola	59°59′N 23°10′E	14.3	3090 ± 60	3350–3190	GrN-19648	−30.26	Beginning of isolation	Eronen *et al.* 1993
	Gundbyträsket, Tenhola			2940 ± 60	3170–3010	GrN-19647	−30.00	End of isolation, Limnaea	Eronen *et al.* 1993
135.	Dotterböleträsket, Tammisaari	60°00′N 23°18′E	8.8	1650 ± 60	1610–1470	GrN-19644	−28.19	Beginning of isolation (?)	Eronen *et al.* 1993
	Dotterböleträsket, Tammisaari			2400 ± 60	2510–2350	GrN-19826	−31.88	End of isolation, Limnaea	Eronen *et al.* 1993
AMS	Dotterböleträsket, Tammisaari			2370 ± 50	2470–2330	GrA-2543	−26.61	Isolation, Limnaea	Eronen *et al.*; this paper

136.	Skogsböleträsket, Tenhola	60°02′N 23°11′E	7.3	2990 ± 70	3250–3070	GrN-17288	−27.87	Beginning of isolation	Eronen *et al.* 1993
	Skogsböleträsket, Tenhola			2420 ± 80	2570–2350	GrN-17289	−30.21	End of isolation, Litorina	Eronen *et al.* 1993
AMS	Skogsböleträsket, Tenhola			1720 ± 50	1690–1570	GrA-2541	−25.54	Isolation, Limnaea	Eronen *et al.*; this paper
180.	Vähäjärvi, Eura	61°10′N 22°12′E	64.2	6500 ± 40	7390–7318	GrN-20902	−26.44	Beginning of isolation	Eronen *et al.* 1995
	Vähäjärvi, Eura			6310 ± 40	7230–7130	GrN-20901	−29.86	End of isolation, Litorina	Eronen *et al.* 1995
181.	Kakkurlammi, Eura	61°58′N 22°14′E	54.3	6600 ± 50	7486–7498	GrN-20904	−22.84	Beginning of isolation	Eronen *et al.* 1995
	Kakkurlammi, Eura			6890 ± 50	7750–7650	GrN-20903	−29.57	End of isolation, Litorina	Eronen *et al.* 1995
182.	Ämmäjärvi, Eura	61°03′N 22°07′E	49.7	5810 ± 30	6658–6590	GrN-21015	−23.30	Beginning of isolation	Eronen *et al.* 1995
	Ämmäjärvi, Eura			5480 ± 40	6310–6210	GrN-21014	−28.65	End of isolation, Litorina	Eronen *et al.* 1995
183.	Urmijärvi, Eura	61°00′N 22°03′E	40.7	4760 ± 40	5530–5434	GrN-21029	−26.03	Beginning of isolation	Eronen *et al.* 1995
	Urmijärvi, Eura			4610 ± 30	5350–5250	GrN-21028	−28.35	End of isolation, Litorina	Eronen *et al.* 1995
184.	Kivijärvi, Laitila	60°58′N 21°54′E	35.6	4370 ± 40	5010–4890	GrN-21019	−25.28	Beginning of isolation	Eronen *et al.* 1995
	Kivijärvi, Laitila			4160 ± 30	4718–4634	GrN-21018	−29.42	End of isolation, Litorina	Eronen *et al.* 1995
185.	Katonajärvi, Lappi T.L.	61°05′N 21°53′E	32.5	4380 ± 40	5030–4910	GrN-21017	−19.96	Beginning of isolation	Eronen *et al.* 1995
	Katonajärvi, Lappi T.L.			4100 ± 30	4650–4550	GrN-21016	−30.22	End of isolation, Litorina	Eronen *et al.* 1995
186.	Vähä Ahojärvi, Kodisjoki	61°02′N 21°45′E	29.4	3700 ± 50	4110–3950	GrN-20906	−24.69	Beginning of isolation	Eronen *et al.* 1995
	Vähä Ahojärvi, Kodisjoki			3600 ± 50	3970–3830	GrN-20905	−26.67	End of isolation, Limnaea	Eronen *et al.* 1995
187.	Rapajärvi, Rauma	61°05′N 21°43′E	23.6	3540 ± 40	3870–3750	GrN-21025	−28.40	Beginning of isolation	Eronen *et al.* 1995
	Rapajärvi, Rauma			3370 ± 40	3650–3550	GrN-21024	−30.20	End of isolation, Limnaea	Eronen *et al.* 1995
188.	Tuitinjärvi, Rauma	61°08′N 21°39′E	20.4	3300 ± 35	3566–3482	GrN-20908	−21.33	Beginning of isolation	Eronen *et al.* 1995
	Tuitinjärvi, Rauma			2970 ± 30	3166–3090	GrN-20907	−29.30	End of isolation, Limnaea	Eronen *et al.* 1995
189.	Tarvolanjärvi, Rauma	61°06′N 21°35′E	18.0	2770 ± 30	2922–2854	GrN-21027	−27.21	Beginning of isolation	Eronen *et al.* 1995
	Tarvolanjärvi, Rauma			2540 ± 30	2662–2582	GrN-21026	−28.18	End of isolation, Limnaea	Eronen *et al.* 1995
190.	Monnanjärvi, Rauma	61°06′N 21°33′E	14.0	2320 ± 40	2390–2270	GrN-21023	−25.24	Beginning of isolation	Eronen *et al.* 1995
	Monnanjärvi, Rauma			2180 ± 40	2210–2110	GrN-21022	−28.76	End of isolation, Limnaea	Eronen *et al.* 1995
191.	Pyytjärvi, Rauma	61°09′N 21°30′E	9.6	1655 ± 40	1598–1510	GrN-20910	−25.69	Beginning of isolation	Eronen *et al.* 1995
	Pyytjärvi, Rauma			1515 ± 40	1450–1370	GrN-20909	−28.71	End of isolation, Limnaea	Eronen *et al.* 1995
192.	Koijärvi, Rauma	61°05′N 21°33′E	8.8	1870 ± 30	1830–1758	GrN-21021	−25.26	Beginning of isolation	Eronen *et al.* 1995
	Koijärvi, Rauma			1645 ± 25	1570–1514	GrN-21020	−27.99	End of isolation, Limnaea	Eronen *et al.* 1995
193.	Olkiluodonjärvi, Eurajoki	61°14′N 21°30′E	1.5	260 ± 50	330–210	GrN-20912	−26.50	Beginning of isolation	Eronen *et al.* 1995
	Olkiluodonjärvi, Eurajoki			410 ± 50	470–370	GrN-20911	−27.55	End of isolation, Present Baltic	Eronen *et al.* 1995
195.	Kaurajärvi, Eura	60°59′N 22°02′E	45.4	5570 ± 40	6410–6322	GrN-21976	−23.31	Isolation, Litorina	Eronen *et al.*; this paper
	Kaurajärvi, Eura			5270 ± 40	6090–5990	GrN-21975	−27.49	Isolation, Litorina	Eronen *et al.*; this paper
196.	Lavajärvi, Lappi T.L.	61°05′N 22°03′E	56.6	6320 ± 50	7250–7130	GrN-21978	−24.99	Isolation, Litorina	Eronen *et al.*; this paper
	Lavajärvi, Lappi T.L.			5920 ± 50	6810–6690	GrN-21977	−28.30	Isolation, Litorina	Eronen *et al.*; this paper
197.	Ruotana, Köyliö	61°10′N 22°23′E	62.9	6680 ± 60	7570–7450	GrN-21980	−26.23	Isolation, Litorina	Eronen *et al.*; this paper
	Ruotana, Köyliö			6190 ± 60	7130–6990	GrN-21979	−30.55	Isolation, Litorina	Eronen *et al.*; this paper
198.	Valkkisjärvi, Laitila	61°01′N 21°47′E	32.4	4210 ± 50	4810–4670	GrN-21982	−22.67	Isolation, Litorina	Eronen *et al.*; this paper
	Valkkisjärvi, Laitila			4040 ± 50	4590–4430	GrN-21981	−26.48	Isolation, Litorina	Eronen *et al.*; this paper

to "glo" and "flada", and evidence for these transitional phases can be detected in the sediment records through high-resolution diatom analyses.

A very common diatom species of the brackish-water lagoons of the Baltic Sea is *Campylodiscus clypeus*. It is so characteristic of the littoral facies of the Litorina Sea that the upper limit of this stage of the Baltic is often called the "Clypeus limit" (e.g. Eronen 1974). *C. clypeus*, along with *Amphora robusta*, was found to occur abundantly in the brackish-water sediments of the present data set. The isolation event is typically represented by a mass occurrence of *Fragilaria* species. Above this isolation level the sediment changes to gyttja and fresh-water diatoms typical of small lakes and ponds become dominant in the assemblages, though brackish-water species may still frequently occur in the sediments above this horizon. For instance, a common species in post-isolation layers is *Nitzschia scalaris*, which seems to indicate a slight saline-water influence in a predominantly fresh-water lake, where the dominant species commonly include *Aulacoseira* spp., *Caloneis* spp., *Eunotia* spp., *Fragilaria* spp., *Pinnularia* spp. and *Tabellaria* spp.

The loss-on-ignition curves along with the radiocarbon dates of isolation horizons are shown in Figs 4 and 5. Because of varying sedimentation rates, the isolation event is represented by a variable thickness of deposit. As a rule, however, the age difference between the lower and upper date for the isolation horizons is not large. The errors at 1σ of the conventional ^{14}C dates for the brackish and fresh-water sediment frequently overlap and a typical time difference rarely exceeds 100–200 years (Table 1). This probably indicates that there is no marked reservoir effect in the coastal waters of the Baltic in Finland, which would otherwise make the brackish-water sediments consistently many hundreds of years older than the overlying lake sediments. It could be, of course, that there is a similar reservoir effect affecting both the brackish and fresh-water sediments, but such an effect has not been demonstrated in any previous studies in the region. In any case, the dates are largely consistent with the stratigraphical sequence and with each other.

Because our study areas are located in the relatively flat coastal zone, early Holocene raised shore levels are not represented here, but can be detected at higher elevation, further inland. The highest basin studied is Lake Vähäjärvi, in the Ol–Py area at 64.2 m a.s.l., but the oldest radiocarbon date (*c.* 8000 BP) is from Lake Stortjärnen, in the Ta–Pe area (Table 1). This date is at least 500 years older than the onset of the Litorina Sea and indicates the time of isolation of the basin from the Ancylus Lake/Mastogloia phase transition. At that time, the rapid regression of the Ancylus Lake was over, water in the Baltic basin was at the same level as that in the open ocean, and therefore, relative sea level was determined by a combination of eustatic rise and isostatic uplift of the land.

Evidence of the Mastogloia phase and the beginning of the transgressive Litorina Sea was found in the sediments of Lake Kirakanjärvi (basin no. 122; Figs 4 and 5) in Perniö. Indicator diatom taxa for the Ancylus Lake phase include *Campylodiscus noricus, Gyrosigma attenuatum* and *G. acuminatum*. Above the Ancylus sediments, taxa characteristic of the isolation phase are found (e.g. *Fragilaria* spp.), but in the same section of sediment (10–15 cm) some diatoms associated with the Ancylus Lake phase persist, but the flora includes also species indicating weak salinity (e.g. *Cocconeis pediculus, Rhoicosphenia curvata*). Above that phase, the representation of *Mastogloia* species (*M. smithii v. amphicephala, M. elliptica*) increases in the flora. This phase in turn gives way in Lake Kirakanjärvi to assemblages typical of shallow-water lagoons which mark the transition to the Litorina Sea. The indicator taxa for this phase include, for example, *Campylodiscus clypeus, Nitzschia scalaris* and *Amphora robusta*. It is quite evident that a minor Litorina transgression affected the basin of Lake Kirakanjärvi following the Mastogloia phase. The lake basin became isolated from the sea by 6800 BP, as is reflected in the diatom assemblages, the fresh-water taxa such as *Eunotia* spp., *Tabellaria* spp. and *Tetracyclus lacustris* replacing brackish-water diatoms.

Reconstruction and interpretation of the shoreline displacement curves

The shoreline displacement diagrams for the Ta–Pe and Ol–Py areas are presented in Figs 6 and 7 (see also Table 1). Because the rate of land uplift varies between these two areas, and even within each study area, the data are projected on separate representative baselines. The baseline for the Ta–Pe area is the 4 mm isobase for the current rate of annual uplift, while that for the Ol–Py area is the 5 mm/yr isobase (as defined in Kääriäinen 1966, cf. also Suutarinen 1983 and Vermeer *et al.* 1988). The altitudes of the lake basins orthogonally to the baseline are corrected using current uplift values, which makes it possible to estimate the effect of crustal tilting. However, the rate of uplift has decreased as a function of time towards the present and therefore the correction applied for tilting slightly underestimates its real magnitude. Attempts to estimate the varying rates of uplift over the past 8000 years had not significantly improved the precision of shoreline displacement curves, because a major slow down in its rate had occurred already before 8000 BP and the study areas are fairly restricted. It is preferable to use present uplift rates for height projections, because they can be measured accurately and possible local anomalies in uplift can be detected.

Two new relative sea-level curves have been constructed in the light of these new data, using both

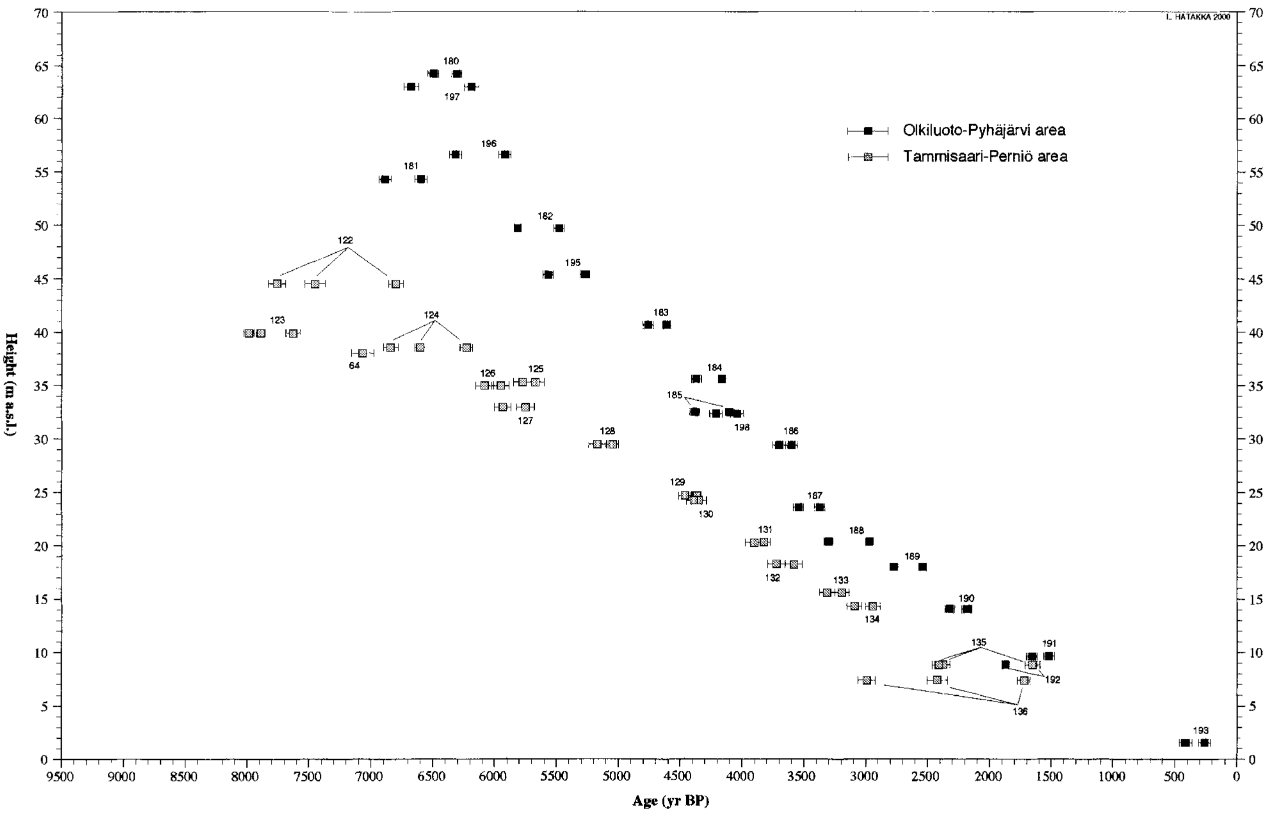

Fig. 6. Data showing relative sea-level lowering in the Olkiluoto–Pyhäjärvi (64, 122–136) and Tammisaari–Perniö (180–198) areas based on conventional radiocarbon dates and elevations without correction for crustal tilting (cf. Fig. 7).

calibrated and non-calibrated radiocarbon dates. They provide somewhat different but complementary information on shore-level displacement (Figs 6, 7). Because of the varying production of ^{14}C in the atmosphere over time, ^{14}C ages (in BP) diverge from calendar years. However, precise calibration curves are available for the Holocene which enable ^{14}C dates to be calibrated to calendar ages (in cal. BP, i.e., calendar years before 1950), including error estimates (van der Plicht 1993; Stuiver *et al.* 1998).

As pointed out by several authors (e.g. Andrews 1986; Pirazzoli 1991), it is extremely important to include a measure of the uncertainties associated with data employed for the reconstruction of sea-level curves. In the data considered here, the altitude uncertainty is of the order of ± 0.5 m in most cases. The uncertainty in the radiocarbon dates is ± 25–80 years (mainly depending on the nature of the sample material). After calibration, the revised uncertainty depends on the precise shape of the appropriate segment of the calibration curve. The conventional radiocarbon dates with error limits are shown in Fig. 4. In Fig. 5, the calibrated dates of the beginning and end of isolation events are combined so that the bar symbols show the range between earliest age of the lower sample and the latest age of the upper sample within calculated error limits.

The shoreline displacement curves of the Ta–Pe and Ol–Py areas indicate a relatively regular rate of uplift and overall relative sea-level lowering with, however, some notable features in parts of the curves. The early (upper) part of the Ta–Pe curve is plateau-like, before changing to a steady declining trend at around 6100 BP (*c.* 7000 cal. BP). Unfortunately there are no sea-level index points old enough to establish whether a corresponding phase of slowed regression also occurred in the Ol–Py area.

In fact a minor transgression seems to have occurred in the Ta–Pe area soon after the beginning of the Litorina Sea stage, as indicated by the diatom evidence in the Lake Kirakanjärvi sequence (44.5 m a.s.l.) in Perniö, in the 3rd Salpausselkä region. As described above, the lake became isolated by 7700 BP, but the diatoms indicate an increase in salinity around 7500 BP. This is likely to reflect a marine incursion and a net sea-level rise, but further detailed studies are needed to confirm this. The final isolation of the lake basin was the emergence from a Litorina lagoon by 6800 BP. Stratigraphical evidence for a Litorina transgression at such a high elevation in Finland has not been reported

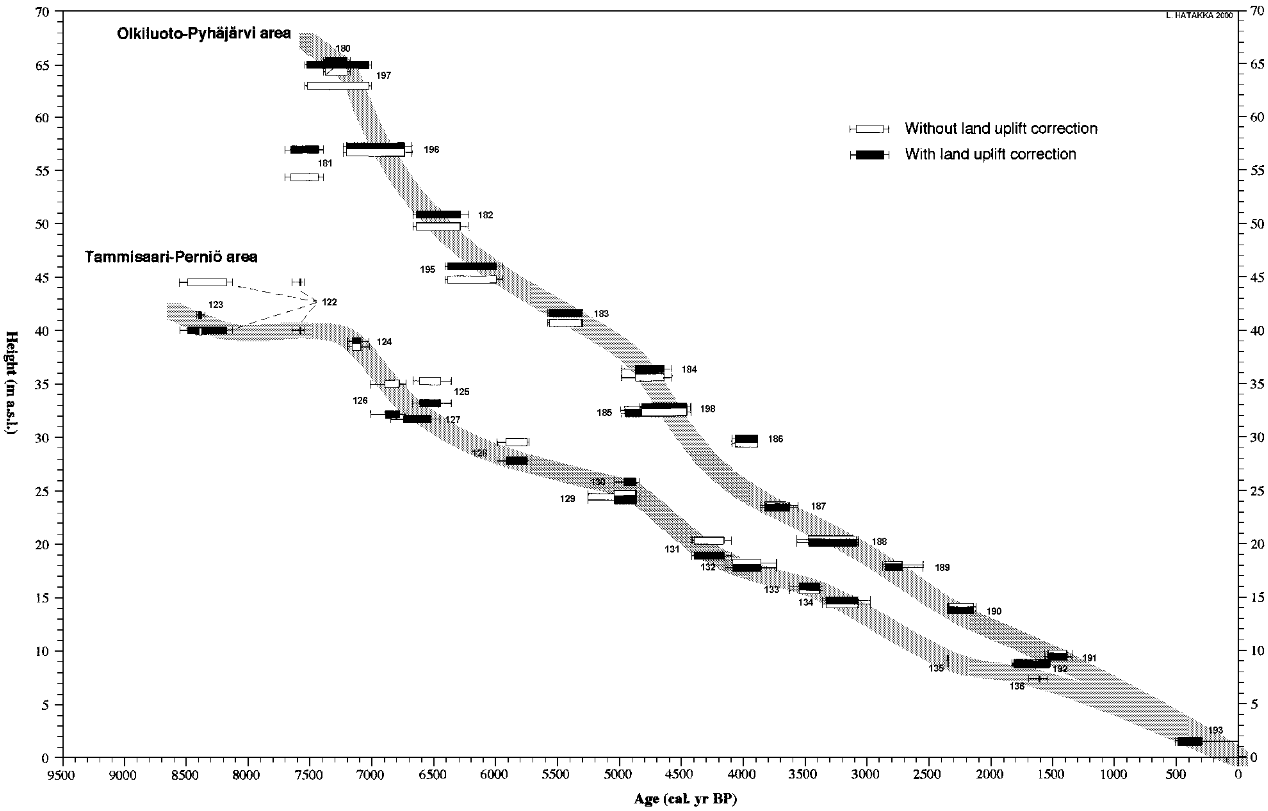

Fig. 7. Curves showing relative sea-level lowering in the Olkiluoto–Pyhäjärvi and Tammisaari–Perniö areas based on calibrated radiocarbon dates. The heights of solid bars a.s.l. are projected to selected baselines to remove the effect of land surface tilting caused by uplift (cf. text).

before. It is consistent with earlier observations by Ristaniemi & Glückert (1988) and Glückert (1991), who have argued that the Litorina transgression reached 42 m a.s.l. (Bastukärr basin) in the 2nd Salpausselkä zone in southwestern Finland (Eronen *et al.* 1993). On the other hand, Hyvärinen (1982, 1999) has not found any indication of a Litorina transgression in his detailed studies on shoreline displacements near Helsinki, which is situated at a lower isobase of land uplift than the sites mentioned above. This discrepancy suggests that there have been marked regional differences in the course of postglacial uplift, but the present data are too sparse to establish the pattern of these differences.

The Litorina transgression resulted from the global eustatic rise in sea level, which affected the shoreline levels all along the coastal areas of Finland, including the Ostrobothnian coast in western Finland. In the latter area, however, the rate of land uplift clearly exceeded the rate of sea-level rise. No evidence for a transgression has been found in the Ol–Py area, which is situated at higher isobases of uplift than the Ta–Pe area, indicating that, in the former area, just as in Ostrobothnia, the rate of sea-level rise during the Litorina Sea stage was lower than isostatic rebound. In that area, and in areas of rapid uplift in general, the rise in global sea level is represented by a decrease in the rate of relative sea-level lowering. Without the influence of eustatic rise, the recorded rate of land emergence would have been much higher in the Mastogloia and early Litorina times than is reflected by the curves.

Even though it is known that melting of the ice sheets generated a substantial rise in global ocean level until 6000–5000 BP (Lambeck *et al.* 1990; Pirazzoli 1991), this component is extremely difficult to determine accurately, because of deglacial geoidal changes (Fjeldskaar 1989, 1994) and global crustal deformations after the end of the last cold stage (Pirazzoli 1991). Removing the influence of eustatic rise would steepen the upper parts of the Ta–Pe and Ol–Py emergence curves, which must be taken into account when estimating the magnitude of crustal deformation in these areas during the past 8000 ^{14}C years.

There appears to be a slight temporary increase in relative sea-level lowering in both curves at *c.* 5000–4500 cal. BP. These effects are unlikely to be artefacts, but there are insufficient data to determine whether a significant increase in the rate of uplift has taken place at those points, or whether they reflect a eustatic effect.

A few additional dates from sites at altitudes between 20 m and 35 m a.s.l. could alter the shape of the curve. The Ta–Pe curve is almost exponential in form, when the effect of the ocean-level rise prior to 6000–5000 BP is taken into account. It can be concluded that land uplift in both these areas has proceeded at a fairly steady rate, but is gradually slowing down (Figs 6 and 7).

Discussion and conclusions

Sediment cores have given no stratigraphic evidence of any transgression after 6000 BP. Most of the isolation horizons provide fairly regular progressively younger dates with descending height a.s.l., but there are some discordant ages, very probably caused by sample specific errors. Bulk sediment dates are commonly affected by hard-water effects (Påsse 1996; Wohlfarth 1996; Björck *et al.* 1996), but it seems that this is not a serious problem in this region, since Precambrian crystalline rocks dominate and no limestones occur in the areas concerned.

In Finland, sediments that have accumulated in small lakes and ponds are therefore generally well suited to radiocarbon dating, but occasionally some as yet unknown factor renders the ages too old by a few hundred years or more. These errors, as well as dates that are thought to be too young, can be checked and at least partly corrected for by using the accelerator mass spectrometry (AMS) method to date seeds or small plant remnants extracted from each isolation level in the sediment sequence. Four such AMS dates were obtained from the present set of samples in the Ta–Pe area from sites with anomalous conventional dates, and these provide results which are more compatible with the smooth shape of the Ta–Pe sea-level curve (cf. Table 1).

Inevitably there are errors in the ^{14}C dates obtained from bulk sediment samples, and small errors can significantly affect the configuration of a sea-level curve. Irregularities can readily be observed when comparing the shoreline curves from different regions of Finland (e.g. Eronen 1983; Matiskainen 1989). It is commonly believed that minor irregularities in these curves, suggesting a lowering of sea level during postglacial time, are mainly due to inaccurate radiocarbon dates. It is possible, however, that they reflect reversals superimposed on the overall lowering trends, but this hypothesis can be tested by employing a larger number of accurate age determinations.

These new Finnish shoreline data, together with the evidence already published, provide no evidence for vertical displacement between bedrock blocks. The results suggest, however, that there may have been irregular differences in the uplift between different regions, and more detailed studies could reveal minor local disturbances of uplift caused by vertical faulting in the course of glacio-isostatic rebound. Marked vertical dislocations have occurred in northern Fennoscandia, but the resulting faults are in the supra-aquatic area, and date back to early postglacial times when uplift was still very rapid (Lagerbäck 1990; Wahlström 1993).

The geophysical data indicate that isostatic recovery will continue yet for several thousands of years, even though there are uncertainties in the calculated rate of residual uplift. Estimates vary from 30 m to 150 m (Ekman 1989), though the most recent calculation suggests an amount of *c.* 90 m for residual uplift (Ekman & Mäkinen 1996). Thus the lowering trends in relative sea levels will prevail in the northern part of the Baltic for a long time, unless greenhouse warming causes a strongly accelerated rise in world ocean level.

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