

## Late glacial multiproxy evidence of vegetation development and environmental change at Solova, southeastern Estonia

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**Abstract.** Reinvestigation of the late glacial Solova (Remmeski) basin, based on plant macrofossil and diatom record, AMS  $^{14}\text{C}$  chronology and sediment composition (loss-on-ignition and magnetic susceptibility data), provided information on vegetation history and palaeoenvironmental and palaeoclimatic changes since the time of the deglaciation of the area around 14 000 cal yr BP. The chronology of the sequence is based on seven AMS dates on terrestrial macrofossils, providing evidence of rapid sedimentation in between 14 000 and 13 500 cal yr BP. Loss-on-ignition data show a clear short-lived warming episode centred to 13 800 cal yr BP, tentatively

correlated with the GI-1c warming of the event stratigraphy of the Last Termination in the North Atlantic region, which suggests that at least parts of the Haanja Heights were ice-free by 14 000 cal yr BP. Macrofossil evidence indicates *Betula nana*–*Dryas octopetala*-dominated open tundra communities with *Saxifraga* on dry ground, and *Carex* sp. and *Juncus* on wet ground at that time. The first evidence of the postglacial presence of tree birch (*Betula pendula*) in Estonia is dated back to 13 500 cal yr BP. However, conifer remains were not found in the late glacial sediment sequence of Solova Bog. The late-Allerød (GI-1a) organic deposits, which are quite typical of other parts of Estonia and indicate general warming, are missing at Solova, most probably due to a hiatus in sedimentation in this very small and shallow upland basin.

**Key words:** plant macrofossils, aquatic macrophytes, diatoms, geochronology, late glacial, deglaciation, Estonia.

### INTRODUCTION

The Late Weichselian deglaciation and the following vegetation formation and succession pattern in the eastern Baltic area have been a topic of interest since Hausen (1913), who initially outlined the ice-marginal positions of the last glaciation in Estonia. The ice-shed line (Karukäpp 2004) amid the glacier flows of the Gulf of Riga (Baltic ice stream complex) and Lake Peipsi–Pskov, starting between the junction of Salpausselkä end moraine arcs in the north (Lundqvist 1987) and running along the Pandivere Upland, and the Otepää and Haanja Heights and the Latgale Upland consecutively in the south, delimits a narrow NW–SE-trending area of a possibly thinner and inactive glacier that may have been more vulnerable to the Bølling–Allerød warming (GI-1; Lowe et al. 2008) between 14 700 and 12 858 cal yr BP. Thus, the southeasternmost part of Estonian territory was probably liberated from ice first, however, the exact time is still poorly known. In addition, this ice-divide might have served as a primary passage for migrating late glacial vegetation.

The aim of this study is to shed light on the much-debated chronology of the onset of deglaciation in the southernmost part of Estonia (e.g. Raukas et al. 2004; Kalm 2006; Rinterknecht et al. 2006). Additionally, we hope to clarify the local vegetation development after

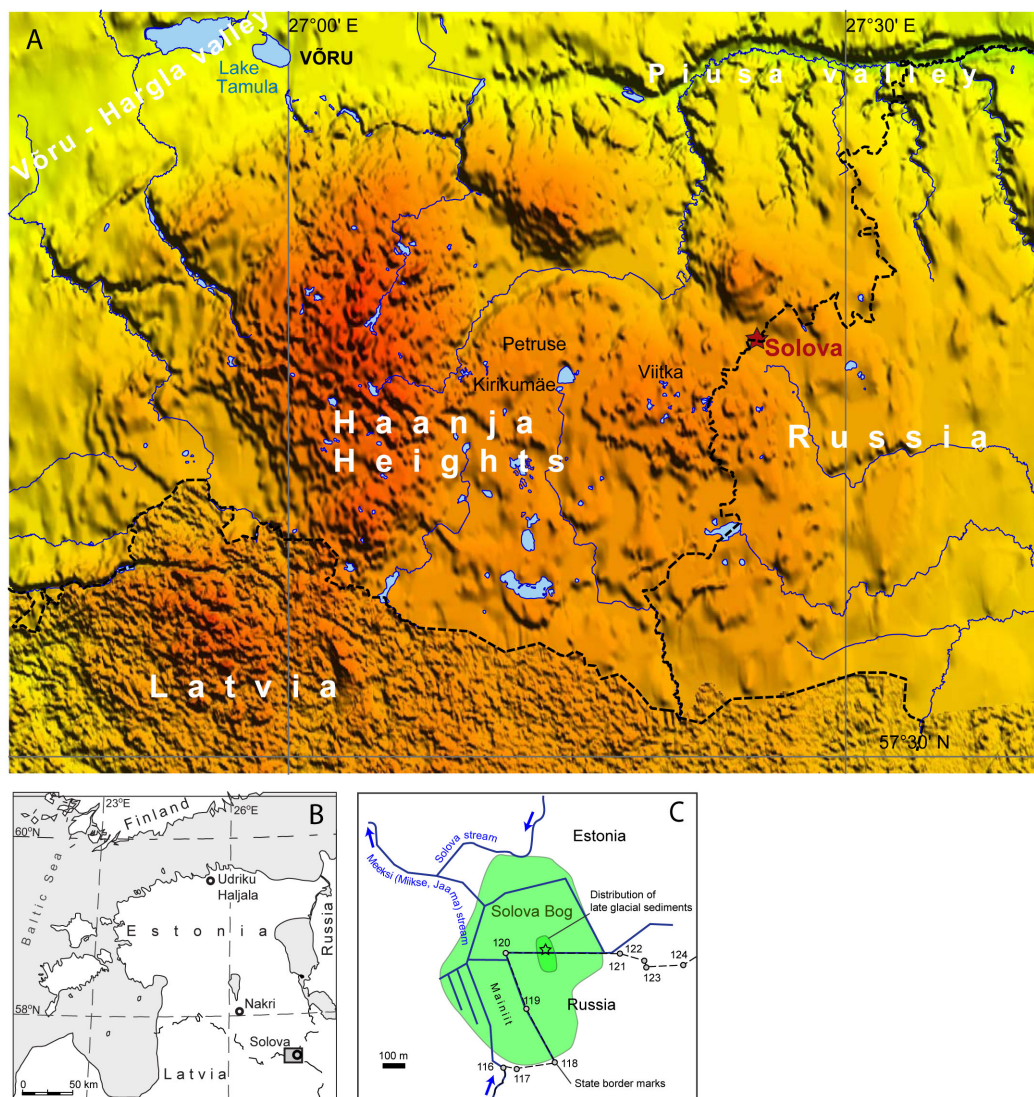
the retreat of ice, which, in turn, may indicate climate change in late glacial time. Fossil plant macrofossils have proven to be an exceptionally reliable tool for reconstructing past local vegetation cover in recently deglaciated areas (Birks 2003). Moreover, the dating of carefully selected terrestrial plant macroremains contributes to the accurate deglaciation chronology. Because of its geographical position at the southern border of Estonia, topographically rather high (168 m a.s.l.) on the Haanja Heights in the ice-shed zone, Solova is probably one of the oldest postglacial sedimentary basins on the current territory of Estonia, which has yielded evidence of late glacial warming. Results of the palaeobotanical study at the Solova site reflect traces of the first vegetation succession and postglacial tree limit dynamics in Estonia after the Last Glacial Maximum (LGM), giving therefore valuable information not only about a local-scale palaeoenvironment, but also about palaeoclimatological shifts in a regional scale. Present results, combined with recent palaeoecological and chronological data from the neighbouring areas (e.g. Stančikaitė et al. 2008; Amon et al. 2009; Heikkilä et al. 2009; Saarse et al. 2009; Amon & Saarse 2010), will contribute to our knowledge about deglaciation of the eastern Baltic area and ice sheet decay in the southeastern sector of the Scandinavian Ice Sheet, as well as about the vegetation succession pattern and timing in these areas.

## STUDY SITE

Solova Bog (Fig. 1; 57°42.024' N; 27°24.915' E) is located in the eastern part of the Haanja Heights (Viitka–Luhamaa hummocky area) at the Estonian–Russian border at an elevation of 168 m a.s.l. The surroundings of the bog reach in places ca 230 m a.s.l. Solova Bog is today a drained and forested wetland some 12 ha in size, however, the basin area containing late glacial deposits is limited only to 75 m in diameter on the Estonian side.

The Solova site, formerly known as Remmeski, is a well-known late glacial reference site for the Estonian stratigraphic scheme (Pirrus & Raukas 1996). There is some confusion, though, in the earlier data as Solova

(12 ha) and Remmeski (118 ha) bogs lie just 5 km apart. The majority of investigations seem to have been conducted at the Solova site but published under the name of Remmeski Bog, and some studies from the Remmeski site under its proper name. Pollen evidence from the ‘real’ Remmeski Bog was first published by Veber et al. (1961) in the frame of investigations of peatland-type sections of upland areas. Pirrus (1969) investigated the late glacial pollen composition of the Solova sequence and speculated on the ‘older than Allerød’ pollen spectra, represented as elevated pine pollen percentages below ‘typical Allerød’ strata (R. Pirrus pers. comm. 1996), and included the site in the stratigraphical scheme of the Estonian late glacial (Pirrus & Raukas 1996). Unfortunately Pirrus (1969)



**Fig. 1.** (A) Location of Solova Bog (marked with a star) in the eastern part of the Haanja Heights elevation model on the border between Estonia and Russia. Location of sites in the vicinity of Solova is marked. (B) Location of the investigated area in southeastern Estonia. (C) The immediate surroundings of Solova Bog.

mixed up the site names and published her results under the name of Remmeski. Radiocarbon dates under the name of Remmeski are available both from the late glacial and Holocene sediment layers (Punning et al. 1971) and have yielded ages around  $12\,750 \pm 145$  cal yr BP for the late glacial part, yet, it is unclear whether these  $^{14}\text{C}$ -dated samples were taken from the Solova or Remmeski site. Despite the confusion with the locality names, the Solova basin remains an important late glacial key site to which we now try to implement AMS  $^{14}\text{C}$  chronology based on terrestrial macrofossils and determine local vegetation development based on plant macrofossil evidence.

## METHODS

In August 2007 a sediment core was taken for AMS  $^{14}\text{C}$  chronology from Solova Bog, using a Russian peat sampler 10 cm in diameter and 1 m long. Sediments down to the sand layer at the bottom were recovered, a maximum sediment thickness of 670 cm was recorded and sediment lithostratigraphy was described. The cores were photographed, carefully packed into 1 m plastic semitubes, wrapped in polyethylene film, labelled and transported to the laboratory. An additional set of replicate cores for macrofossil analysis was taken from exactly the same place as the previous coring point in September 2008.

The chronology of the sediments was established using seven AMS radiocarbon dates on terrestrial plant macrofossils. The radiocarbon dates were converted to calibrated median ages using the IntCal04 calibration dataset (Reimer et al. 2004) and the OxCal 4.0 program (Bronk Ramsey 2001). All ages mentioned in the text refer to calendar years before present (cal yr BP; 0 = AD 1950).

The organic matter (OM) content was measured as loss on ignition after heating dried samples at  $550^\circ\text{C}$  for 4 h (Heiri et al. 2001). Magnetic susceptibility (MS) was measured from the whole length of the core on the sediment surface at 1 cm resolution using a Bartington Instruments Ltd. high-resolution surface scanning sensor MS2E. Measurements were performed from the cleaned sediment surface covered with a thin plastic film. Results of MS values are expressed in SI units.

Plant macrofossil analysis was carried out in contiguous 4 cm intervals from the sediment core at depths between 660 and 480 cm. The uniform subsample size (approximately  $300\text{ cm}^3$ ) was determined by displacement of water in a measuring cylinder. The preparation for plant macrofossil analysis followed conventional procedures (Birks 2001). Material retained on sieves was examined using stereo- and light microscopes. The material identified was mostly plant macro-

fossils (seeds, catkin scales, endocarps, etc.) but also remains of different aquatic animals and moss fragments. Moss analysis was performed for the core depth interval 657–557 cm. Several specimens were identified to genus level due to poor preservation or identification problems. The group *Drepanocladus sensu lato* sums up different *Drepanocladus* and *Warnstorfia* species. *Sphagnum* was also united into one group due to lack or detachment of stem leaves. Plant macrofossil and moss atlases (e.g. Aalto 1970; Abramov & Volkova 1998; Ingerpuu & Vellak 1998; Cappers et al. 2006) were used for identification. In addition, seed collections and herbaria of the Department of Environmental Archaeology, National Museum of Denmark, and W. Szafer Institute of Botany, Polish Academy of Sciences, were employed. The moss nomenclature follows that of Ingerpuu & Vellak (1998). Plant macrofossil zonation follows constrained incremental sum of squares (CONISS) cluster analysis (Grimm 2007).

For diatom analysis freeze-dried sediment was weighed and the samples were digested in 30%  $\text{H}_2\text{O}_2$  until organic material was oxidized. A few drops of 10% HCl were added to remove carbonates and thereafter the fine mineral particles were removed by repeated decantation. Commercially available divinylbenzene microscopic markers were added to determine diatom concentration. A few drops of the cleaned subsample were dried onto the cover glass and mounted on slides using Naphrax medium and analysed for microfossils under a Zeiss Imager microscope at  $\times 1000$  magnification, using oil immersion and differential interference contrast optics.

## RESULTS

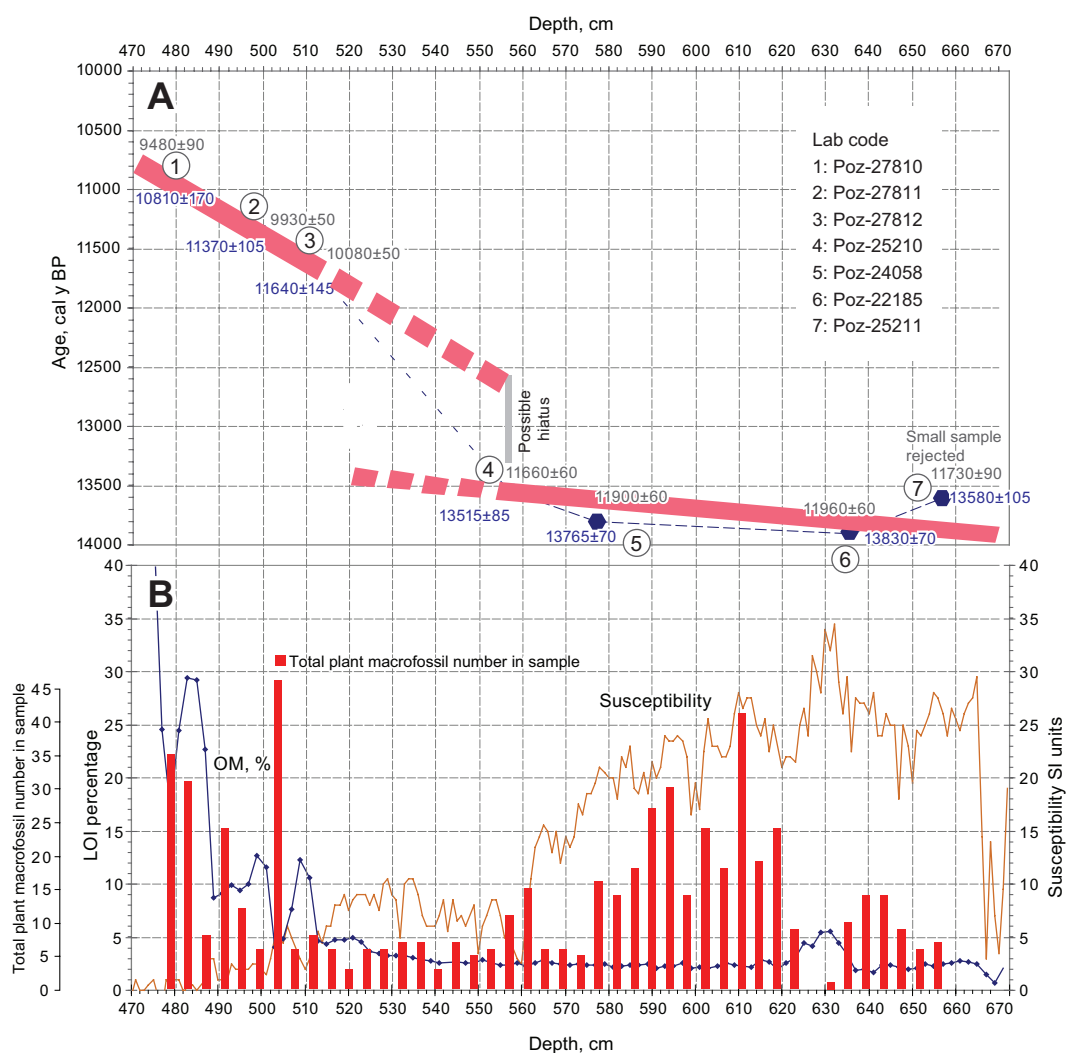
### Lithostratigraphy and chronology

Only the lower part of the sediment core, from 670 to 480 cm, was studied. Details of the sediment lithostratigraphy are presented in Table 1. The basal part of the section consists of 180 cm thick minerogenic sediments that are overlain by gyttja and peat deposits. Coarse-grained clastic sediments, namely medium and fine-grained sands, which constitute the lowermost sediments of the sequence, have overall low OM values. The MS values are very low at the base of the core but increase notably further up in the sequence. These sediments are interpreted as having accumulated immediately after the deglaciation. A distinct shift in the sediment grain size and MS values occurs at a depth of 560 cm. Above the 487 cm level, minerogenic sediments are overlain by OM-rich lacustrine gyttja sediments.

The chronology of the core is based on seven AMS dates on terrestrial macrofossils (Fig. 2A). An age/depth plot shows that the lower part of the sediment section

**Table 1.** Lithostratigraphic description of the sediment sequence at the Solova site

Depth from the surface, cm	Sediment description
450–487	Gyttja, homogeneous, olive to green; OM = 17–82%; MS = 0–1 ( $\times 10^{-5}$ ) SI
487–516	Silt with diffuse organic matter and plant fragments; light greenish-grey, distinctly darker more organic-rich layers at 493–495, 498–502 and 508–512 cm; OM = 4.0–12.6%; MS = 0–6 ( $\times 10^{-5}$ ) SI
516–560	Clayey silt/silty clay, grey; OM = 2.3–4.0%; MS = 3–11 ( $\times 10^{-5}$ ) SI
560–624	Silt, beige; OM = 2.1–2.9%; MS = 12–28 ( $\times 10^{-5}$ ) SI
624–637	Clayey silt, massive, light beige; OM = 3.4–5.6%; MS = 24–35 ( $\times 10^{-5}$ ) SI
637–666	Silt with sandy interlayers, laminated, grey; OM = 1.7–2.8%; MS = 18–30 ( $\times 10^{-5}$ ) SI
666–672	Sand, medium to fine-grained, brownish-grey; OM = 0.7–2.1%; MS = 3–19 ( $\times 10^{-5}$ ) SI



**Fig. 2.** (A) The age–depth model of late glacial sediments of Solova Bog with the possible hiatus indicated. Chronology is based on  $^{14}\text{C}$  dates and calibrated ages BP (median age with  $2\sigma$  error margins). The bottommost sample No. 7 was rejected from the age–depth model due to unexpected results from a very small sample (less than  $0.2\text{ }\mu\text{g C}$ ). (B) Graph showing LOI percentages (solid line with diamond markers), magnetic susceptibility in  $10^{-5}$  SI units (solid line) and total plant macrofossil number (bars) along the depth scale.

(core depth 630–560 cm) deposited around 13 800–13 500 cal yr BP and suggests a high sedimentation rate (over 0.2 cm/yr). The lowermost AMS  $^{14}\text{C}$  date produced a ‘younger than expected’ age, possibly due to a very small quantity of dated material and was rejected. The dating at a depth of 512 cm yielded an age of  $11\,640 \pm 145$  cal yr BP, i.e. the Pleistocene–Holocene boundary. The other dates from the upper portion of the core were consistent and suggested a uniform but apparently slower sedimentation rate compared to the lower portion of the core. The age/depth curve implies a possible hiatus in the sediment record (Fig. 2). Considering the distinct change in lithostratigraphic parameters, the hiatus could be located at a depth of 560 cm, on the boundary between silt and clay sediments. In that case a gap of about 1000 years in sedimentation persists, i.e. between ca 13 500 and ca 12 500 cal yr BP, suggesting that the climatic events of the Last Termination GI-1a and GI-b and the initial part of GS-1 (Lowe et al. 2008) are partly absent from the Solova sediment record.

### Plant macrofossils

The macrofossil record of the Solova basin is divided into six macrofossil assemblage zones (SMZ) in accordance with CONISS cluster analyses. Results of the macrofossil concentrations are given in Fig. 3.

In the lowermost zone (SMZ-1, 660–636 cm, fine-grained sand/silt) the abundance of plant macrofossils is low; only few scattered remains of arctic plant taxa are present. In SMZ-2 (636–576 cm, clayey silt) the majority of taxa are similar to those in SMZ-1; however, the abundance of macrofossils increases up to 40 finds per sample. The finds are mainly leaves of *Dryas* sp., but in addition, the remains of *Betula nana*, *Juncus* and *Potamogeton filiformis* are noted.

In SMZ-3 (576–557 cm, clayey silt/silt) the abundance and richness of plant species declines compared to the previous zone. In SMZ-4 (557–534 cm, silty clay/silt) macroremains of *Juncus* disappear and the number of *Daphnia ephippia* increases. At the lower boundary of the zone a remain of *Betula pendula* was recorded. Around the core depth of 530 cm, in SMZ-5 (534–497 cm, silt with plant fragments), a clear shift in macroremain composition is observed (Fig. 3). Arctic species such as *Dryas octopetala* and *Betula nana* almost disappear above this level and the macrofossil assemblage is dominated by aquatic and telmatic plant remains (Potamogetonaceae, *Ranunculus* sect. *Batrachium*). The abundance of Characeae oospores increases simultaneously. The uppermost zone SMZ-6 (497–479 cm, silt with plant material/gyttja) contains macroremains of tree-birch, *Alnus* and *Phragmites*, and suggests a shift in the vegetation towards an

assemblage of a much warmer climate. Moreover, in SMZ-6 the burst of limnic animal remains occurs – hundreds of *Daphnia ephippia* and *Cristatella mucedo* statoblasts were identified.

### Diatoms

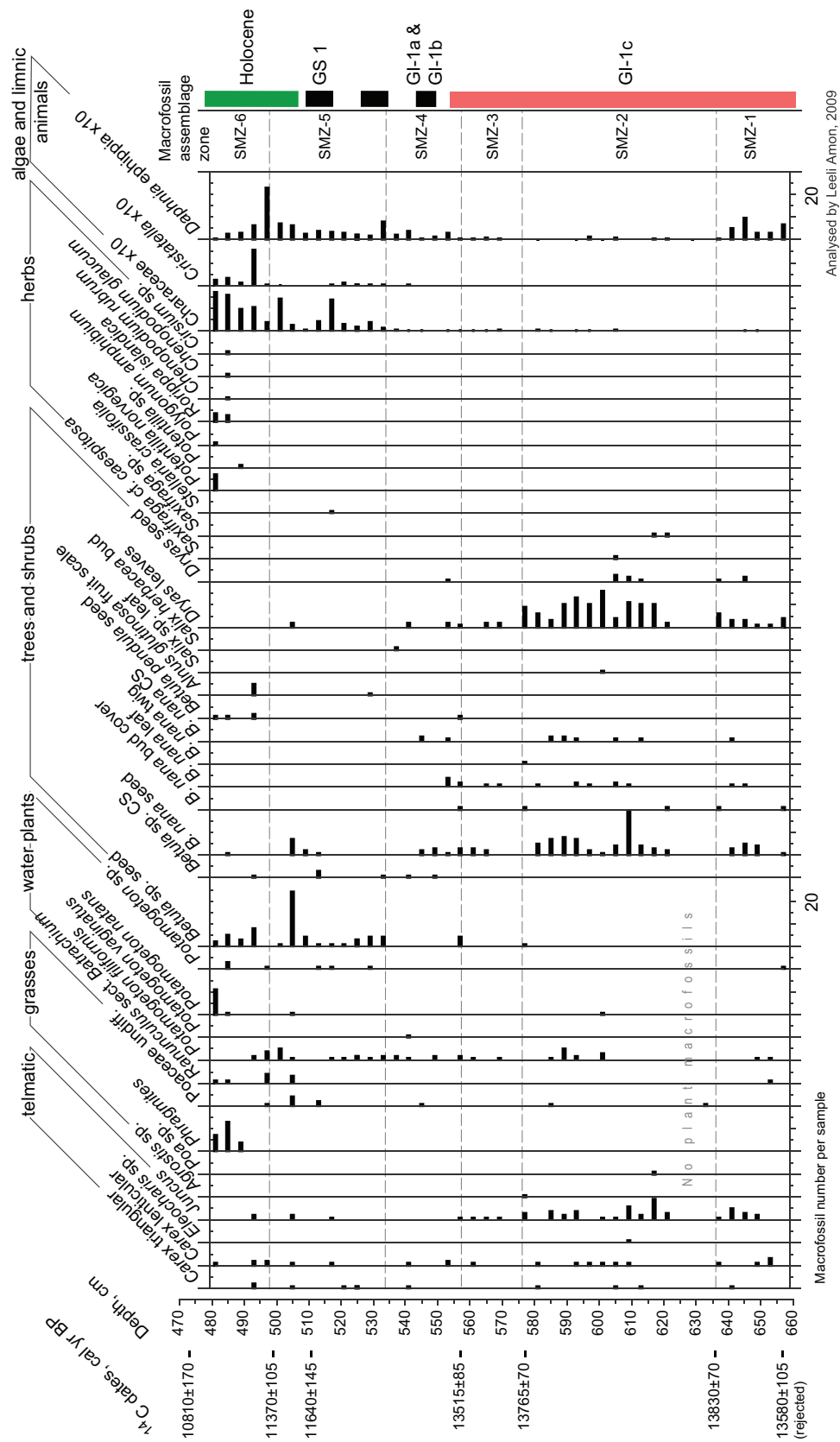
Most of the lower part of the studied sediment sequence was devoid of diatoms (Fig. 4). This can be ascribed to a high accumulation rate of mineral particles, possibly causing considerable dispersion of diatoms. Diatom scarcity can also be attributed to poor preservation through dissolution of diatom frustules. From a core depth of 500 cm benthic diatoms appear, notably small-sized fragilarioid diatoms. From a depth of 485 cm, the diatom composition becomes more diverse. The first planktonic diatoms (e.g. *Cyclotella comensis* and *C. comta*) appear, *Fragilaria* species decline and are replaced by epipellic taxa such as *Navicula radiosa* and *N. diluviana*, as well as by epiphytic/epipsammic *Achnanthes minutissima*.

## DISCUSSION

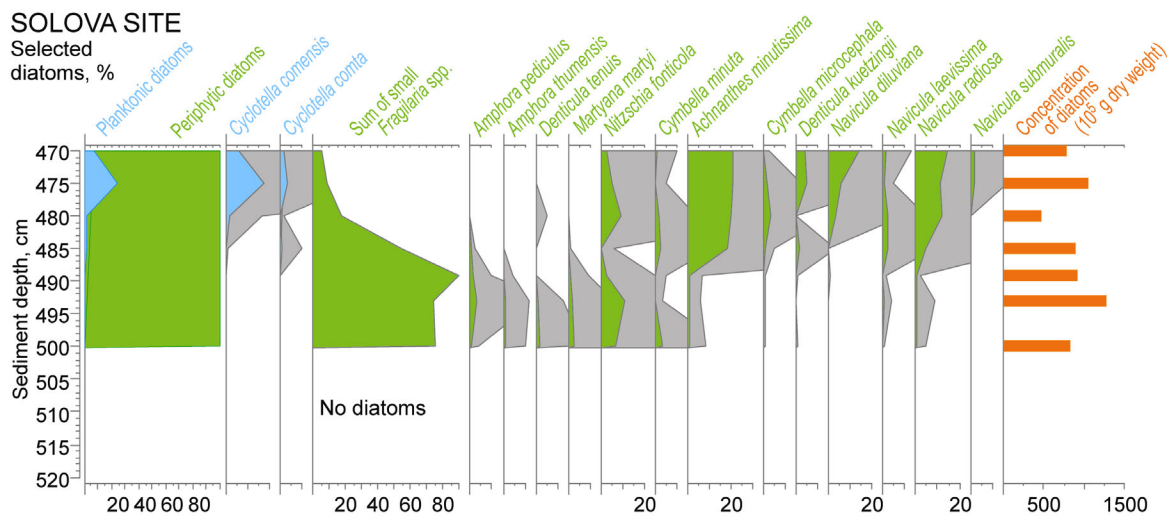
### The development of terrestrial environment at Solova

The terrestrial vegetation record at Solova begins at approximately 14 000 cal yr BP (Fig. 3). Correlation of the studied time interval with the ice core data suggests that the lower portion of the sedimentary record at Solova (660–557 cm) can be associated with the GI-1c event (Lowe et al. 2008), a 650-year-long slightly warmer episode of the late glacial period. Species richness in the lowermost part of the sediment core is low. The main species found, *Betula nana* and *Dryas octopetala*, are typical pioneer plant species of the late glacial period, which provide information about the local terrestrial biota, suggesting rather severe climate conditions up to 13 500 cal yr BP (660–556 cm). In addition, *Juncus* and *Carex* remains indicate wet surroundings of the locality. Between core depths of 635 and 627 cm around 13 800 cal yr BP adjacent samples are practically empty of plant macrofossils, except for one grass seed. This coincides with a slight rise in the OM content of sediment (Fig. 2). High MS values at the same level are likely indicative of catchment erosion and fast material (including paramagnetic particles) inwash into the basin, possibly supplemented with higher soil organics input, rather than increased autochthonous organic production, which otherwise would seem to be the most likely explanation for higher OM content. Lack of plant macrofossils in this small section may be related to local amelioration shown by more organic-





**Fig. 3.** Macrofossil diagram for Solova Bog (number of specimens per sample) plotted against age (cal yr BP) and correlated with schematic Greenland ice core chronology (Lowe et al. 2008). CS, catkin scale.



**Fig. 4.** Diatom stratigraphy of the main taxa as relative abundance (%), proportion of plankton and periphyton (%) and diatom concentration ( $10^5$  g dry weight) from the Solova Bog sediment sequence. The grey curves show  $10\times$  exaggeration of the percentage values.

rich sediments, but the deterioration of preservation conditions does not allow us to draw further conclusions about local vegetation development, based on indicators of terrestrial environment. After ca 13 800 cal yr BP the terrestrial vegetation reflected by plant macrofossils consists of arctic species and suggests open tundra landscape, proved by the presence of light-demanding species like *Dryas* and *Saxifraga*. Remains of telmatic species such as *Juncus*, *Eleocharis* and *Carex* are indicative of wetland around the lake. From ca 13 750 cal yr BP the abundance of plant macroremains falls considerably but the species composition is more or less the same, i.e. *Betula nana*, *Dryas* leaves and *Juncus* seeds. The first macroremain find of the real tree-birch *Betula pendula* from the Solova sediments (559–555 cm) may refer to a short-time climate warming at 13 500 cal yr BP, just before the suggested hiatus. However, from this point onwards the use of the exact temporal scale is hampered by a possible sedimentary hiatus (Fig. 2).

The total plant macrofossil concentration is the lowest at the 556–510 cm level (Fig. 2), suggesting colder Younger Dryas conditions. A distinct rise in the abundance of plant macrofossils occurs at a depth of 510 cm at the Younger Dryas–Preboreal boundary. A series of samples consist of *Betula* sp. seeds – wingless or decayed birch seeds. These seeds could be either dwarf or tree birch seeds.

The transition to the Holocene at ca 11 650 cal yr BP (Lowe et al. 2008) is revealed by a principal change in plant species composition. Both, tree species (*Betula pendula*, *Alnus glutinosa*) and herbs are reflecting

ameliorated environmental conditions. In the section analysed the variety of herbs is the largest in the early Holocene. The presence of *Phragmites* suggests the proximity of wetland, as does the occurrence of *Carex*, *Juncus* and *Polygonum amphibium* seeds. On the other hand, the terrestrial herb community enlists *Potentilla norvegica*, which nowadays is largely present in sandy and gravelly grounds all over the Scandinavian Peninsula. *Rorippa islandica* is indicative of wetter clayey ground, showing as well unstable open surfaces around the sampling site.

#### The development of aquatic environment

The initial assemblage of aquatic organisms in the Solova basin at ca 14 000 cal yr BP consisted of two plant species, *Potamogeton filiformis* and *Ranunculus* sect. *Batrachium*. Both are cold-tolerant species, usually associated with relatively turbid waters, and are likely to imply an intensive mineral soil inwash from the catchment (Välranta 2006). *Daphnia* ehippia were present at the start of postglacial sedimentation. *Daphnia* spp. often colonized newly created water bodies (Sarmaja-Korjonen 2003; Sarmaja-Korjonen et al. 2006), e.g., they were the first immigrants in the late glacial Lake Kråkenes in western Norway (Birks 2000). In Solova, 13 800 cal yr BP, however, the amount of aquatic remains declined, except for a small cluster of *Potamogeton filiformis* endocarps and the appearance of a cosmopolitan pondweed species, *Potamogeton natans*. The shortage of limnic macroremains around a core depth of 630 cm rather suggests

an event of rapid inwash of autochthonous OM and considerable minerogenic contribution (high MS values) to sediment matrix, which caused a pronounced dilution effect of fossil remains. Alternatively, a harsh climate with long-lasting ice cover, when only a narrow marginal belt of open water existed in the lake during the short summer season, precluded the development of the aquatic biota populations in the basin.

From approximately 13 600 cal yr BP (570 cm), the occurrence of *P. filiformis* endocarps in the samples is almost continuous albeit in small numbers. From the same time onwards, also *Daphnia* remains are constantly present in sediment and show a rising trend.

The only find of *Potamogeton* cf. *vaginatus*, a nowadays rare Nordic pondweed species that prefers deep brackish-water conditions, comes from 540 cm depth (suggested age around 12 300 cal yr BP). At the same time *Cristatella mucedo* statoblasts appear in the sediment record and the abundance of Characeae oospores rises considerably. This change in the community of water organisms precedes the rising trend in the concentration of terrestrial macrofossil remains. A similar pattern – the increased abundance of algae, although microscopic ones, followed by higher concentrations of plant remains – is observed at the Pleistocene–Holocene boundary at Nakri (Amon et al. 2009). Above the 530 cm level (around 12 000 cal yr BP) oospores are present in all samples, numbering between 4 and 160 per sample. However, the counts of oospores are relatively small compared to other studies where thousands of Characeae oospores were recovered in smaller sample volumes (e.g. Birks 2000), suggesting a limited amount of fructifying algae in the given conditions.

Temporarily, during the end of the Younger Dryas or the onset of the Holocene, *Ranunculus* sect. *Batrachium* reappears, indicating rise in water temperature (Birks 2000). The population of limnic organisms and Characeae is rich also in the Holocene part of the studied section. Temperature-sensitive *Cristatella mucedo* (Økland & Økland 2000) displays a maximum at 11 300 cal yr BP (490 cm), confirming mild conditions in water environment at that time.

Small-sized *Fragilaria* spp. are often associated with high environmental instability. They reproduce quickly and tolerate shorter environmental oscillations, which makes them very competitive in unstable limnological conditions (e.g. Haworth 1976; Korhola & Weckström 2004). Thus, the peak of the pioneer forms of *Fragilaria* spp. in the Solova sediment profile may temporarily be linked with the onset of Holocene rapid climatic change (Fig. 4).

The appearance and increase of planktonic diatoms, namely *Cyclotella* spp. (485 cm), in the sediment record may be related to milder and stabilized climate in the

Early Holocene. During warmer years the ice cover is less extensive, and hence the growing season is lengthened and planktonic diatoms may be relatively more abundant. Marked increase in the abundance of planktonic diatom assemblages is apparent in the 150-year-long palaeoclimate record from northern Lapland lakes, reflecting recent climatic warming (Sorvari & Korhola 1998). However, the overall diatom composition and dominance of littoral diatoms indicate a very shallow hard-water lake for the Solova basin.

## Mosses

Only a few studies of moss fragments as a possible palaeoecological proxy are available from Scandinavia (e.g. Bennike & Hedenäs 1995; Jonsgård & Birks 1995) and a few from Estonia, whereas the latter deal mostly with *Sphagnum* species and mires/wetlands (e.g. Sillasoo et al. 2007). However, the richness of moss species and well-characterized recent ecological amplitude should encourage the use of moss analysis for (late glacial) palaeoenvironmental studies (Janssens 1990). Only a limited number of bryophyte species were identified from the sediments of Solova Bog. The presence and absence of species is presented on the stratigraphic scale in Fig. 5.

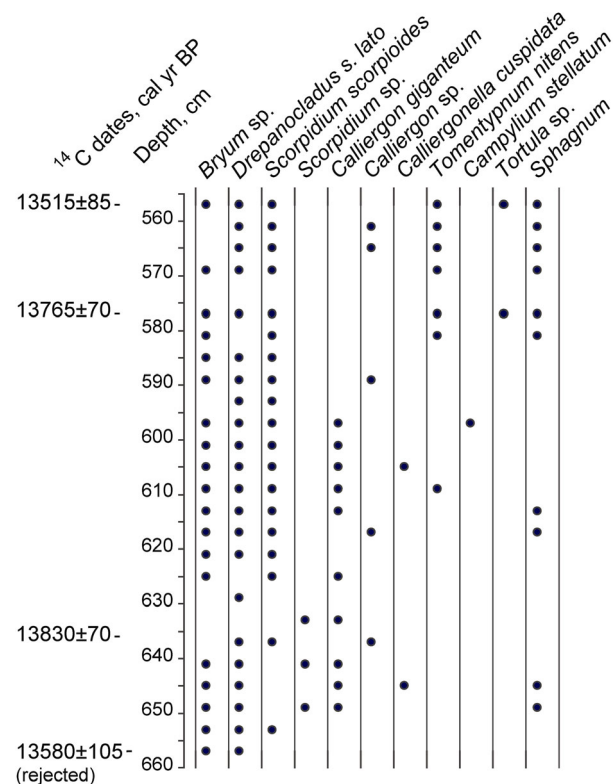


Fig. 5. Moss species presence/absence in the lower section of the sediment core from Solova Bog.



Moss fragments identified (from 657–557 cm) are represented mostly by widespread aquatic mosses in the lower part of the sediment section, with a succession of fen and even more terrestrial species upwards in the core. *Scorpidium scorpioides* is present throughout the section; this species has also been found since the Saalian period in Denmark (Odgaard 1981). *Calliergon giganteum* was identified in the lower half of the analysed sediment sequence. Today *C. giganteum* is a circumpolar moss species that prefers water-rich environment (Koponen et al. 1995). *Tomentypnum nitens* appears first in the middle of the section and is constantly present in the upper part of the sequence. *Tomentypnum nitens* is found also in the earliest Holocene sediments of Svalbard and is interpreted as a species occurring in mineral-rich or calcareous wetland habitats (Bennike & Hedenäs 1995). Another fen-indicative species is *Campylium stellatum*, the only occurrence of which is in the middle of the sediment core (approximately 13 850 cal yr BP). The presence of *Tortula* in the upper part of the sequence clearly shows the proximity of a shore or a higher ground as all recent species from this genus that grow in Estonia prefer terrestrial and even dry habitats.

#### Pollen evidence of environmental change

The majority of lakes and bogs on the Haanja Heights are relatively young, formed predominantly in the Preboreal (e.g. Punning et al. 1995; Niinemets & Saarse 2006) due to glaciokarstic processes. Only in a few of them, such as Lake Kirikumäe and Solova (Remmeski) Bog, postglacial sediments started to accumulate in the late glacial period (Pirrus 1969; Saarse & Rajamäe 1997).

The Solova sediment core has been studied for pollen content (Pirrus 1969; Sarv 1983) and is considered as one of the type sections for Estonian late glacial stratigraphy (Pirrus & Raukas 1996). Solova pollen spectra represent typical late glacial vegetation composition comprising grasses, sedges and dwarf birch, however, a specific section with a high percentage of tree pollen occurs below layers of the Younger Dryas flora. This feature – high amounts (over 80%) of pine pollen, is characteristic of the Allerød in earlier published pollen stratigraphies of Estonian late glacial sediments, which we in the light of recent data from the adjacent area (Amon et al. 2009) can attribute to the latest part of the Allerød (GI-1a). Unfortunately, there were no valid radiocarbon ages tied to pollen stratigraphy and therefore it is questionable to correlate recent macrofossil data and the pollen diagram of Pirrus (1969).

Features of vegetation development similar to that at Solova are seen in the Lake Kirikumäe pollen record

from the late glacial period (Saarse & Rajamäe 1997). The oldest radiocarbon dates from Lake Kirikumäe are from the Preboreal, but pollen data extend into the late glacial so-called Allerød (GI-1a or GI-1b according to Lowe et al. 2008), indicating that a certain area in the eastern Haanja Heights has been ice-free since 14 000 cal yr BP, i.e. after the Older Dryas.

#### Reconstruction of terrestrial vegetation cover in southeastern Estonia

Tree species migration and tree line dynamics in the late glacial period (Willis & van Andel 2004; Kullman 2008) and in the Holocene (e.g. Giesecke & Bennett 2004) have deserved attention in modern palaeoecology due to their relationship with the contemporary climatic and environmental conditions. In areas affected by the ice front recession of the LGM, climate ameliorations/deteriorations are important factors for re-introduction of species into the local environment. In respective studies plant macrofossil data, although with similar limitations, may be more favourable than pollen evidence due to low pollen productivity in lower temperature conditions and possible resedimentation of pollen from earlier time periods. Pollen spectra from closely spaced sites (Solova and Lake Kirikumäe) indicate high pollen percentages for trees (birch, 60%) in the early late glacial. Macrofossil evidence, though, implies no tree cover at Solova at that time. We do not have chronological control over pollen evidence at Solova, but we argue that the pre-Allerød higher tree pollen values show non-local production and may coincide with the higher OM content at 13 800 cal yr BP at Solova (Fig. 2), thus suggesting the reflection of the GI-1c warm episode in the recently ice-freed environment of SE Estonia. The herb/shrub composition (*Artemisia*, Chenopodiaceae, grasses and sedges) in pollen evidence of the same time differs/complements the macrofossil data displaying a grass–sedge–shrub tundra community.

At the Solova site macrofossils of mostly shrubs are found, the most common of them being dwarf birch (*Betula nana*), which is present throughout the late glacial to the Holocene. *Betula nana* is also commonly mentioned in other studies of late glacial sediments of the region (e.g. Wohlfarth et al. 2002) and has been reported to have grown near the ice margin even during the LGM (Binney et al. 2009). *Salix* remains at Solova occurred in small quantities; however, willows (probably of shrub growth forms) are not rare in the Estonian late glacial vegetation, as shown by evidence from the Udriku (Amon & Saarse 2010) and Nakri (Amon et al. 2009) localities. Different *Salix* species have also been recognized in

various regions in Eurasia from before the LGM to the present (Binney et al. 2009).

Tree-birch species, of which *Betula pendula* was identified in Solova sediments, are more indicative of climate warming. The first occurrence of the tree-birch is approximately 13 500 cal yr BP. At the adjacent (70 km N) Nakri site (Amon et al. 2009) the first occurrence of the tree-birch was dated around 13 400 cal yr BP, that is ca 100 years later than in Solova. In modern Fennoscandia birch forms the forest–tundra transition zone (Hustich 1983) and it seems to be the case in the eastern Baltic late glacial as well.

It is noteworthy that only 130 km to the southeast of Solova, at Lake Kurjanovas in eastern Latvia (Heikkilä et al. 2009), the onset of forest and its composition were largely different. Logically, the southern regions further away from the ice edge became forested earlier: the closure of the *Pinus* forest at 14 000 cal yr BP is proposed as the first pine macrofossils date from that time. The southern boundary of the present circumpolar Arctic climatic zone is the northern extent of the closed boreal forests. There is no clear boreal forest boundary but a transition from south to north represented by the following sequence: closed forest → forest with patches of tundra → tundra with patches of forest → tundra (Callaghan et al. 2004). The described transition zone is relatively narrow (30–150 km) and quite comparable with the late glacial situation. Tree-birch seeds were found in Lake Kurjanovas sediments, followed by spruce macrofossil around 13 000 cal yr BP, indicating closed forest conditions just 130 km southeast of Solova (Heikkilä et al. 2009). Even further south, in Lithuania, conifer macrofossils were described from the transition from the Younger Dryas to the Holocene (Stančikaite et al. 2008). In principle, spruce may have survived cold episodes of the late glacial period in the northern refugia, as indicated by spruce megafossils in the Scandes Mountains, dated to ca 12 900 cal yr BP and pine dated to 13 500 cal yr BP, i.e. the Allerød warm period (Kullman 2008). In the eastern Baltic area nunatak refugia are hard to imagine and spruce as a continental species (Giesecke et al. 2008) most probably overglaciated somewhere in Belarus (Giesecke & Bennett 2004). East of Estonia, in Karelia, conifers were indicated by pollen only in the Preboreal period (Wohlfarth et al. 2002). Neither at Solova nor at Nakri the macrofossil assemblages (L. Amon et al., unpublished data) did contain conifer remains; even tree-birch remains were scarce, suggesting a rather distinct late glacial forest line somewhere in northern Latvia. The only trace of conifers so far discovered from Estonian late glacial deposits is a few pine stomata (dated back to ca 13 300 cal yr BP) from Lake Nakri, suggesting only rare pine trees in the area.

## Deglaciation history

Quaternary glaciations and their geomorphologic features have been studied in many aspects in Estonia (Raukas et al. 2004). Classically, five ice-marginal zones have been distinguished. There are several age estimations from mineral samples (using thermoluminescence, optically stimulated luminescence and cosmogenic  $^{10}\text{Be}$  methods) as well from OM ( $^{14}\text{C}$  datings) and clay varve chronology (Hang 2003; Kalm 2006; Rinterknecht et al. 2006). According to interpretation of pollen diagrams, the retreat of the ice margin from the Haanja zone began in the Bølling period and the area of Estonia was finally ice-free in the second half of the Allerød chronozone (Pirrus & Raukas 1996). The age of the Haanja stage is based on controversial age estimations from the Raunis locality in north-central Latvia (Zelčs & Markots 2004). Radiocarbon age estimations from several sites near Solova, such as Viitka intertill organic layer (Fig. 1) dated to  $12\,925 \pm 65$  and  $13\,020 \pm 100$  cal yr BP (calibrated from Punning et al. 1967, 1981), and similar layers at Petrusse with ages around  $14\,850 \pm 350$  and  $13\,940 \pm 150$  cal yr BP (calibrated from Punning et al. 1981) imply ice advance as late as 13 000 cal yr BP if we consider those organic layers, especially at Viitka, as interstadials, which afterwards were covered with stadial (Haanja) till. The present results and age estimations from Solova, where open sedimentary conditions have persisted since 14 000 cal yr BP, support the conclusion that the unexpectedly young ages from the Viitka intertill organic layer can be regarded as caused by probable slope processes or glaciokarstic slumping where the younger organic material was buried under till as suggested earlier (Karukäpp et al. 1992; Raukas et al. 2004). The ages of Petrusse layers, on the other hand, are not contradicting with the ages from Solova, referring to possible ice re-advance around 14 000 cal yr BP (GI-1d), formerly associated with the Older Dryas cooling. Although other evidence around the Haanja Heights, namely accumulation of varved clays in the Tamula basin from 14 700 BP (Kalm 2006) would place Petrusse organic layers into the same category of probable slope-processes/glaciokarstic slumping origin as the Viitka layer. The elevated location of Solova (about 170 m a.s.l.) on the Haanja Heights and possibly close to the ice margin of the Haanja stage allows us to use its basal ages as references for deglaciation of the area. On the other hand, Kalm (2006) associated the Haanja stage with minimum ages around 14 700 cal yr BP, taking the corrected age of Tamula clay varve chronology into account (Sandgren et al. 1997). A probable ice-shed line (Karukäpp 2004), better visible along the Haanja and Otepää Heights (less in Pandivere and Vooremaa) and

appearing again at Salpausselkä (Lundqvist 1987; Boulton et al. 2009), delimits a narrow NW–SE-trending zone of a possibly thinner and partly inactive glacier that may have been more vulnerable to the Bølling–Allerød warming, as nearly simultaneous breakdown of ice (not necessarily meaning that the area became totally ice-free but rather suggesting that openings in the ice cap developed) occurred in southern Estonia (at Solova prior to 14 000 cal yr BP) and in the north near Haljala and Udriku ca 13 800 cal yr BP (Saarse et al. 2009; Amon & Saarse 2010). However, vegetation development differed markedly along that gradient.

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## Hilisjääaegne taimeistiku ajalugu ja keskkonnamuutus Solova soo näitel

Leeli Amon, Atko Heinsalu ja Siim Veski

Solova soo hilisjääaegsete setete uurimine näitas, et järveline vabaveeline settebassein moodustus umbes 14 000 kalendriaastat tagasi. On täheldatav lühiajaline soojenemine umbes 13 800 aastat tagasi, taimeistik Solova ümbruses oli tundrilmeline vaevakase-drüüase kooslus. Esimene Eesti pinnal leitud pärastjääaegne (puu)kase makrojäänus on umbes 13 500 aastat vana.