International Field Symposium of the INQUA Peribaltic Working Group
Tartu, September 13–17, 2009

EXTENT AND TIMING OF THE WEICHSELIAN GLACIATION SOUTHEAST OF THE BALTIC SEA

University of Tartu,
Institute of Ecology and Earth Sciences
Tartu 2009

Organisers:

University of Tartu, Institute of Ecology and Earth Sciences, Department of Geology.
University of Latvia, Faculty of Geography and Earth Sciences.
INQUA Peribaltic Working Group (INQUA TERPRO Commission).
ESTQUA – Estonian National Committee of INQUA.

Organising Commitee:

Volli Kalm, Maris Rattas, Vitalijs Zelčs, Tiit Hang, Alar Rosentau, Kadri Sohar, Liina Laumets, Katrin Lasberg, Marko Kohv.

Recommended reference to this publication:

Kalm, V., Laumets, L. & Hang,T. (eds.), 2009. Extent and timing of the Weichselian Glaciation southeast of the Baltic Sea: Abstracts & Guidebook. The INQUA Peribaltic Working Group Field Symposium in southern Estonia and northern Latvia, September 13–17, 2009. Tartu Ülikooli Kirjastus, Tartu, 112 pages.

Sponsored by:

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Joint Stock Company "Seda" Skolas laukums 1, Seda, Valka District, LV-4728, Latvia,, phone: +371 64722113, fax: +371 64781101

Editors: Volli Kalm, Liina Laumets & Tiit Hang.

The front cover photography "Late Weichselian till on glaciofluvial deposits, Kambja endmoraine" (photo by Maris Rattas). Graphical elaboration by Katrin Lasberg.

ISBN 978–9949–19–218–2

Tartu University Press
www.tyk.ee

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LATE-GLACIAL MULTIPROXY EVIDENCE OF VEGETATION DEVELOPMENT AND CLIMATE CHANGE AT SOLOVA, SOUTHERN ESTONIA

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A 1.5 m thick sediment sequence from a very small Solova mire (57°42.024 N, 27°24.915 E) – formerly known as late-glacial Remmeski site (Pirrus 1969) at an elevation of ca 165 m a.s.l was revisited and provided information on palaeoenvironmental and palaeoclimatic changes since the time of the deglaciation of the area around 14,000 cal yr BP based on plant macrofossil and diatom record, AMS ^{14}C chronology and sediment composition, namely loss-on-ignition and magnetic susceptibility data. A chronology of the site is based on 7 AMS dates on terrestrial macrofossils gives evidence of rapid sedimentation in between 14,000–13,500 cal yr BP. Loss-on-ignition data shows a clear short-lived warming centred to 13,900 cal yr BP tentatively correlated with GI-1c warming of the event stratigraphy of the Last Termination in the North Atlantic region (Lowe et al. 2008), which suggests that at least parts of the Haanja Heights were ice free by then and at least some of the inter-morainic dates of organic material, such as Viitka nearby, must be reconsidered in terms of ice re-advances. Macrofossil evidence points to *Betula nana – Dryas octopetala* dominated open tundra communities with *Saxifraga* on dry ground and different *Carex* sp. and *Juncus* on wet ground. First evidence of tree birch comes around 13,500 cal yr BP, which is coeval with the evidence from Nakri site, another well-dated late-glacial site ca 80 km north-west from Solova. The end-Allerød GI-1a organic deposits quite typical to other parts of Estonia (Saarse et al 2009) indicating general warming at 13,300–12,900 are missing at Solova most probably due to interrupted sedimentation in this very small and shallow upland basin.

Acknowledgements: this research was supported by ETF Grant 7029.

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ENVIRONMENT AND CLIMATE EVOLUTION IN THE NORTH OF RUSSIA DURING THE QUATERNARY

Lyudmila Nikolaevna ANDREICHEVA, Yulia Vladimirovna GOLUBEVA, Tatyana Ivanovna MARCHENKO-VAGAPOVA

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Recently the problem of global warming, connected with the rapid development of industry, is actively discussed all over the World. However, the common opinion concerning the global warming is absent in the scientific community.

In the recent years the palaeoclimatic researche, allowing reconstructing the environmental and climatic dynamics during the Quaternary with the purpose of the forecasting their development in the future, has especially intensively developed. The reconstruction of the environment during the previous interglacial epochs, which according to our data have been several degrees warmer than the current period of the Holocene, is of a great importance.

The mean annual and July temperatures and the amount of the precipitation during the warm and cold intervals of the interglacials were estimated by V.A. Klimanov's information-statistical technique (Klimanov 1976). The vegetation history and character of the climatic changes in the North of Russia within the Timano-Pechoro-Vychegodsky region have been defined.

The palynological data suggest the warmer and moderately damper climate in comparison with the current climate during the Early Pleistocene Vishera Interglacial. An accurately expressed climatic optimum on the spore-pollen diagrams is absent. The mean July temperature reached to18–20^{o}C, which is 6^{o} higher of the current temperature in the North, and 2–4^{o} higher from that in the South of the region. The amount of annual precipitation during the warm interval was 350–400 mm which decreased to to 150–175 mm during the cold interval.

During the first Middle Pleistocene Chirvino Interglacial the region was characterized by warmer climate than at present. In some sections two climatic optima are established. During the first optimum the climate was warmer and damper, than at present, and forests of *Abies* and *Picea* with *Betula, Pinus* and the broadleaved trees grew. During the second optimum the climate was drier and cooler and the broadleaved trees and exotic species were present as admixture. At the final phase of the Interglacial the dominance of tundra was typical. In the north of the region the climate that was 2–4°C warmer, than at present, and the mean July temperature was 14–16^{o}C. In the south of the area the mean July temperature varied between 16 and 18^{o}C, which is 1–2^{o}C higher of the current. Amount of the precipitations in the north of the region was 350 mm and in the south about 255 mm during the warm interval. The amount of precipitation During the cold interval the amount of precipitation did not exceed 50–75 mm in the area.

During the second Middle Pleistocene Rodionovo Interglacial the vegetation was more xerophilous in comparison with the Chirvino Interglacial. There are two climatic optima within the Iinterglacial. The conditions of the earlier optimum of the Rodionovo Interglacial were warmer than the climatic optimum of the Holocene. Dark coniferous forest of the southern taiga type grew, with exotic and broadleaved trees (Duryagina & Konovalenko 1993, Andreicheva & Marchenko-Vagapova 2004). In the North of the region the mean July temperature was about 16^{o}C, which is 4^{o}C warmer of the current. In the southern part the mean July temperature was 16–18^{o}C which is 1–2^{o}C warmer than at present. During the warm interval the amount of precipitations reached 300–400 mm in the north of the region and 350–400 mm in the south. During the cold interval it decreased twice, down to 175–200 mm.

The first Late Pleistocene Sula Interglacial is characterized by a considerable warming. One climatic optimum has been determined during which broadleaved species grew and two maxima of coniferous trees occurred. The mean July temperature in the North of the region was 14–16^{o}C that is 3^{o}C warmer than the current temperature. The amount of precipitations during the warm interval varies between 350 and 400 mm, and during the cold interval it wass essentially less – 150–175 mm.

At the Byzovaya time within the Late Pleistocene, the climate was cooler, than that of the previous interglacials. In this connection the question about the rank of the Byzovaya warming remains open. In our opinion, it is Megainterstadial (Andreicheva & Duryagina 2005). The mean July temperatures in the region were approximately the same as now in the northern part (10–14^{o}C) of the region. In the south of the region the temperature was 2–6^{o}C colder, than it is at present. The precipitation during the warm interval both in the north and in the south of the region was 350–400 mm, and during the cold interval it decreased to 200 mm on the average.

During the Holocene as well as during the previous Interglacial epoch, the repeated vegetation changes occurred. Within the Holocene five climatic periods (Preboreal, Boreal, Atlantic, Subboreal and Subatlantic) were characterized by L.D. Nikiforova (Nikiforova 1980). The Holocene climatic optimum conditions developed at the end of the Atlantic period. This is proven by the distribution of the most thermophilic tree species and the maximum amount of precipitation and highest temperature. Boreal and Subboreal temperature maximums have of secondary subordinate value. The warming of the climate at the end of the XX century and currently is of the local character in comparison of the general tendency.

Thus, according to our data, the climate in the North of Russia is colder in comparison with the previous Interglacial epoch of the Pleistocene. Holocene is considered to be the next Interglacial. Its palaeogeography practically does not differ from the last interglacial epoch. The periods of the cold climates in the Earth history, which repeatedly occurred, were accompanied by the formation of the continental glacial shield. Naturally and rhythmically the glacial epoch were replaced by the interglacial. This provides the basis for the assumption of a gradual cooling in the future that does not exclude the possibility of the glacier. So contrary to the forecasts, the global cold, instead of the global warming is coming.

Acknowledgements: this research was supported by the Grant from the Division for Earth Sciences of RAS (№ 12).

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A PALAEORECONSTRUCTION OF GLACIATION IN DRENTHE, THE NETHERLANDS; WITH A REFERENCE TO NW EUROPEAN GLACIAL LANDSCAPES

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Incomplete knowledge of the Quaternary history of Drenthe provides a missing link in landscape planning, the protection of geoheritage values and nature- and landscape development. That is the main conclusion of recent palaeogeographical studies in Drenthe, the Netherlands. Two studies will be presented.

1. Case-study brook valley system of the Dwingelderstroom, Drenthe, the Netherlands. A Quaternary geological study as a base for geoconservation and nature development.
2. Ice sheets, ice streams and surge glaciers in Drenthe. A glaciological interpretation of Saalian till as a base for the actual landscape and ecohydrological processes.

1. Digital terrain analyses suppose in the central part of Drenthe an ancient river pattern, in the western part connected to an overridden push-moraine and opposite to the southern direction of the recent stream. The supposed change of the course of the river may be caused by a number of processes: e.g. down-cutting by changes of the base-level, proglacial reactivation of faults as a result of forebulging or postglacial uplift in an area with salt domes after the retreat of the land ice. Results of a field study are discussed in relation to results of regional analyses of deep subsurface geological structures, glacial subsurface features of till deposits and geomorphological features and topography. The results of the study include a reconstruction of the post-glacial palaeolandscape.

Effects of uplift or up doming in the postglacial period after the Saalian glaciation had no direct influence on subsequent landscape forming processes. On the other hand, pro-glacial reactivation of older faults cannot be excluded, because of a lineament formed by faults, the boundary of Elsterian deposits and Saalian till, which are also reflected in the actual topography of the surface. A second conclusion is that glaciers followed existing pro-glacial valley systems.

2. Studies on the local level, as mentioned above, contributes to a better understanding of the kind of ice-movement at the regional and European scale.

A Glacial Model (GM) of ice flow in Drenthe and NW Europe will be presented. The GM is based on glacial footprints (e.g. glacial features; till analyses: the basement; thickness; sources of clasts; clay minerals) with a reference to recent glaciology studies and other concepts of the Saalian glaciation. Recent glacial landscape features are to explain by different ice-sheets, ice-streams and surge glaciers related to geological structures (e.g. position of salt domes; hard rock barriers and older glacial features and palaeolandscapes)

Finally, conclusions are related to recent plans for nature- and landscape development in Drenthe. More knowledge about the genesis of landscapes contributes to better protection of (subsurface) geoheritage values, biodiversity and to provide a better instrument for a sustainable environmental planning.

PLANT MACROFOSSIL AND POLLEN EVIDENCE OF ECOLOGICAL CHANGES IN THE LITTORINA LAGOON SEQUENCE AT THE PRIEDAINE, SOUTHERN COAST OF THE GULF OF RIGA

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The last palaeoecological investigations of the Holocene lagoonal sediments at the southern coast of the Gulf of Riga were made in 60-s of last century (Aboltina-Presnikova 1960, Berzin 1967).

Palynological investigations of sediments in core No 20 in the locality of Priedaine Stone Age settlement (Kalnina at al. 2009) shows organic sediments accumulation in lagoon start at the Middle Atlantic time about 6500 yrs ago.

Landscape was relatively open during that time – area mainly covered with sedge and grass, but in the dry places broad-leaved and alder forest was distributed. Accumulation conditions have been changed in Priedaine lagoon. Gyttja with sand and silt layer and presence of salty water diatom species indicates frequent seawater inflow in the lagoonal lake. This feature perhaps was related to sea water level fluctuations and storms. Lagoonal lake gradually becomes fill in and gyttja layer were covered by sedge-sphagnum peat during Subatlantic Time about 2800 years ago. Ruderal and cultivated land herb pollen indicates man presence and activities in the area since the beginning of organic sediment accumulation period. Especially intensive accumulation has been during Middle Atlantic (AT2) and Late Atlantic (AT3) but later it decreased.

Data of plant macro remain analysis demonstrates aquatic plant presence in whole section (2.9 – 0.17 m).

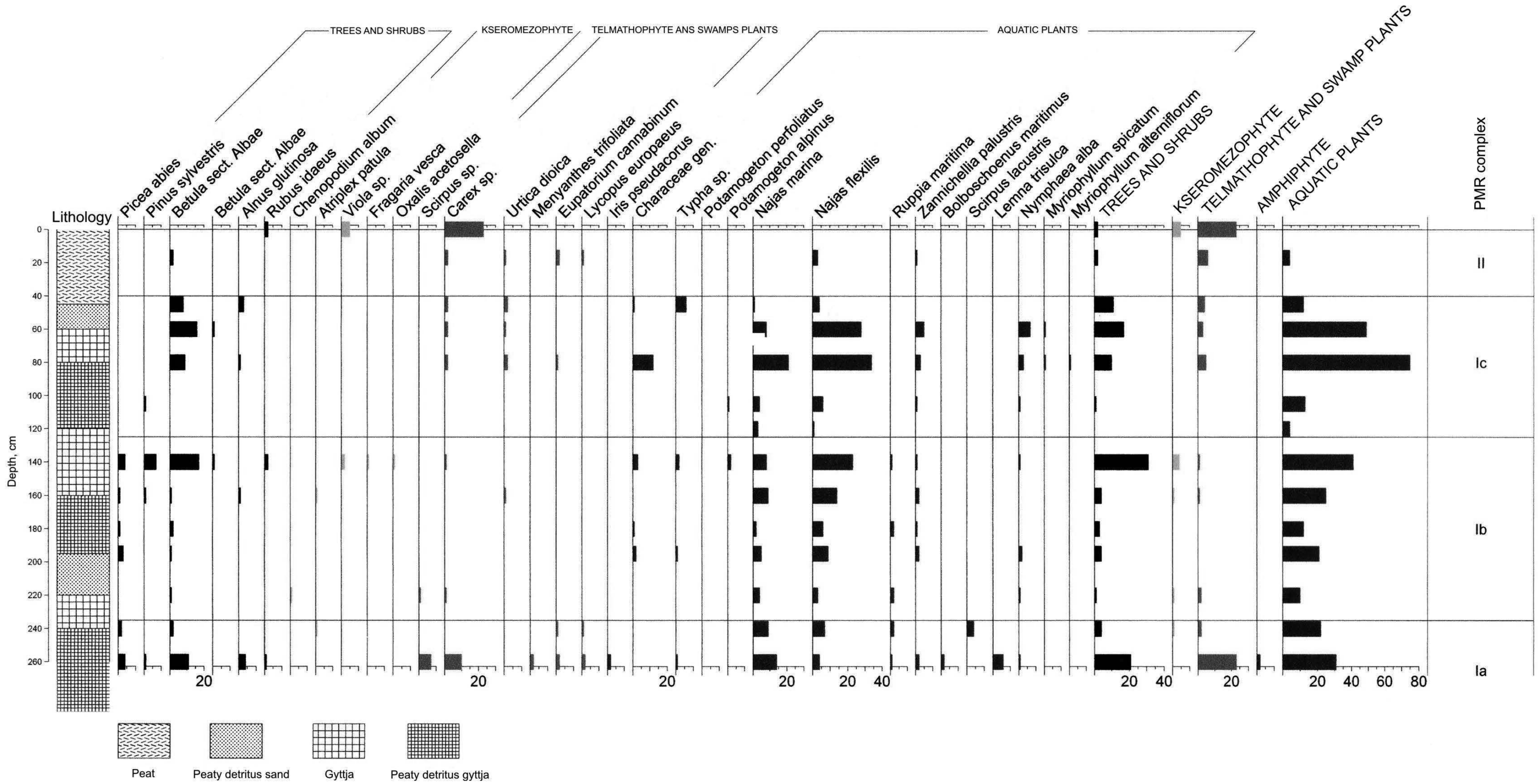
Four macro remain complexes are divided:

Ia – dominance of terrestrial plants, depth interval of zone corresponds to AT2 pollen zone in diagram.

Ib – *Ruppia maritima* presence among aquatic plants (beginning and middle part of AT3). Well-preserved seeds of aquatic plant Widgeon Grass (*Ruppia maritima*) occur in almost whole interval 2.9 – 1.4 m. Plant is distributed in salty seawater in depth to 1.4 m. In Latvia it usually grows together with Soft Hornwort or Tropical Hornwort (*Ceratophyllum submersum*) and Horned Pondweed (*Zannichellia palustris*) in the protected area of Randu meadows located at Vidzeme coastal area of the Gulf of Riga. Seeds of these plants occur as well in this depth interval (*Ceratophyllum submersum* up to 1.6 m). *Ruppia maritima* and *Ceratophyllum submersum* haven't been found above depth 1.4 m. This fact probably indicates that seawater didn't flow into lagoon anymore during next phase of lagoon development and water became less salty and unsuitable for growing of these plants.

Ic – aquatic plants *Nymphaea alba, Zannichellia palustris, Najas marina, N. flexilis* and increase of mire plant remains (end of AT3, SB).

II – dominance of terrestrial plant remains (end of SB, SA). Sharp decrease of aquatic plant species and plant remains as well increase of mire, wet meadow and other terrestrial plants can be observed in the upper part of section (0.6 – 0.17 m). That point on rapid paludification of area and overgrowing of water basin. Only two aquatic plant remains have been found in depth interval 0.45 – 0.17 m, but above this interval just mire plants.



Plant macro remain diagram from Priedaine site (Cerina 2009)

Data of both pollen and macro remain analysis indicate Priedaine lagoon development since the middle part of the Atlantic Time (AT2) until end of it (AT3). Lagoon became as lake during the Subboreal (SB) and continues to fill in and mire has been developed in the area of the former Littorina Sea lagoon during the Subatlantic Time.

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RECONSTRUCTION OF PALAEOVEGETATION AND SEDIMENTATION CONDITIONS IN THE NORTHEASTERN AREA OF ANCIENT BURTNIEKS LAKE

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Pollen and plant macro remain analyses carried out in the frame of the international multidisciplinary project “Archaeological complex Zvejnieki” during last years (2005 – 2008) from sediments of Burtnieks Lake Northern part improve data obtained until year 2005 (Kalnina L., Medne L.). Palaeobotanical investigations have been carried out with aim to reconstruct palaeovegetation development and sediment formation conditions of ancient Burtnieks Lake North-Eastern area.

Authors have acquired fieldwork methods, including geological coring, sounding, visual estimation, and description of the deposits and sampling. Pollen analysis has been carried out at the Laboratory of the Quaternary Environment Faculty of Geography and Earth Sciences University of Latvia. Computer programs TILIA and TGView have been used for pollen diagram construction. Most of the fieldwork was done during the archaeological excavation carried out with aim to find out borders of archaeological site Braukšas II.

Pollen data show vegetation dynamics since Younger Dryas (DR3) until nowadays (Table 1) in the ancient lake coastal zone.

Obtained pollen composition suggests that landscape was relatively open during Younger Dryas with *Betula nana* and other representatives of the Subarctic flora, which disappeared in the very beginning of the Holocene, during the first part of the Preboreal. Trees, shrubs, and birch-pine forests occupied study area and open areas decreased. Pollen spectra reflect natural development of the regional vegetation during the Preboreal, with small number of anthropogenic indicators. Composition of tree pollen shows significant differences in vegetation composition even in small distance intervals during the Boreal. That feature point on local character and intensive activities of ancient man. Pine and pine-birch forest distribution near Pantene and all surrounding area of the Burtnieks Lake is characteristic for the Boreal. Although, differences in herb pollen proportion in total pollen composition reflect human impact on the landscape and vegetation. Similar characteristics reflect also plant macro remain data which indicate presence and activities of ancient man in Pantene area. Pollen data show appearance and distribution of alder and hazel in the forest composition. Fluctuation of pollen curves in diagram can reflect both climatic changes and human impact on vegetation.

Fluctuating curves of broadleaved tree pollen in diagrams, together with increasing amount of ruderal plants, charcoal dust and cultivated plant pollen occurrence during the Atlantic Time, indicate that ancient man not only gathered but also started farming activities.

Sharp decrease of birch and increase in spruce pollen during the Subboreal indicates changes of climate and human impact on nature. Pollen composition from sediment sequences shows that pine and birch are dominating in study area, while number of ruderal plants considerably decreased during the Subatlantic time.

Pollen and plant macro remain composition indicate that gyttja started to form in different times in different places of the ancient Burtnieks Lake: at the end of the Preboreal in the vicinity of core Cerini-2007 and at the second part of Boreal (BO2) in the Pantene Mire and Ruja Mire. Composition of plant macro-remains in peat and data obtained from pollen analyses shows that fen peat in Pantene Mire and Rūja Mire started to form in the end of the Atlantic Time, but in the surrounding territory of Pantene Drumlin (core Cerini-2007) in the beginning of the Boreal.

Table 1. Comparison of local and regional pollen zones in cores Ceriņi-2007, Pantene/BraukšasI-2006, Rūja-2005 and mean pollen diagram of Northern part of Vidzeme

Cores Regional zones	Pantene/ BrauksasI-2006	Cerini-2007	Ruja-2005	Mean pollen diagram of Northern part of Vidzeme (Seglins 2002)
SA3	Be-Al-Pn	Pn-Be	Pn-Be-Al	Pn-Be
SA2			Pc-Pn-Be	Pc
SA1			Pn-Al	Be-Al
SB2	Pc-Pn	Pc-Al-Ti	Pc	Pc
SB1	Al-Be		Be-Al-Pn	Pc-Pn-Ti
AT3	Ti-Car-Pc	Al-Ul-Ti	Ti-Carp-Pc	Q-Ul-Ti-Al
AT2	–	–	Be-Al-Cor	
AT1	Ul-Al-Cor	Pn-Ti	Al-Cor-Ul	Ti-Cor
BO2	Be-Pn-Al	Cor-Be	Be-Al-Pn	Be-Ul
BO1	Pn	Pn-Be	Pn	Pn
PB2	Be-Pn	Be-Poac	Be-Pn	Be-Pn
PB1	Pn-Be		–	–
DR3	Be nana-herbs	–	–	–

MIDDLE WEICHSELIAN PALAEOENVIRONMENTAL RECORD FROM THE SVIRKANCIAI OUTCROP (NW LITHUANIA)

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Investigations of the Middle Weichselian sequences in the Baltic region are of great importance for determination of extension of the Middle Weichselian glaciation (especially during OIS 4) and palaeogeography of the OIS 3. Knowledge on climatostratigraphic events of the Middle Weichselian (Pleni-Weichselian, Pleniglacial) in the Baltic region so far is very scarce if to compare this time span with the Early Weichselian. An extensive area of distribution of palaeolacustrine sediments (sand, clay, silt with humus and interlayers of peat) c.a. 77 km^2, occurring under the relief forming Upper Weichselian till, was determined in vicinities of Venta settlement (North West Lithuania) in the course of geological mapping.

The Svirkanciai outcrop was investigated in 2008, and the sequence reflects presence of nonglacial palaeoenvironments before the Late Weichselian ice advance. The Svirkanciai outcrop (56º18'05" N, 22º53'00" E) is situated on the left abrupt bank of the Virvyte River, the tributary of the Venta River, 6 km to south from Vieksniai town in Venta Regional Park. The outcrop is 15 m height and stretches up to 50 m by the river loop.

Two different genesis and age sediments complexes compose the sequence. The lower part (up to 7.7 m above the river water level) deposited by silt and very fine grained sand. In the upper part of the outcrop Upper Nemunas (Late Weichselian) till overlies. The till is composed by very hard basal till and massive ablation till. The interlayer of fine-grained sand is in between.

The pollen and spores analysis was performed for 43 samples underlying the till: 20 samples of silt, 2 samples of sandy silt and 21 sample of very find grained sand.

The frequencies of pollen are low. Only few pollen grains were identified in sandy sediments at interval 0.25–1.1 m. Several pollen grains were slightly corroded. Colonies of green algae *Pediastrum* were noted. Pre-Quaternary spores as well as *Pinus Haploxylon* were also identified.

Pollen of trees, shrubs and dwarf shrubs constitute 65–90% of the total pollen sum. *Pinus* (pine) pollen (with 60%) is prevailing among the tree pollen , *Betula* (birch) pollen reaches up to 25%, *Picea* (spruce) appears with 10%. A few pollen of *Salix*, *Juniperus, Ephedra* and *Ericales* were noted. *Betula nana* pollen comprises some 8% of the pollen flora.

A rich herb pollen flora was noted e.g. *Apiaceae*, Asteraceae, *Artemisia, Cyperaceae*, *Chenopodiaceae, Poaceae, Rumex*, *Rosaceae, Ranunculaceae*. *Cyperaceae* with 20% predominate. A few pollen of *Helianthemum* were identified as well as spores of *Botrychium* and *Selaginella selaginoides*. Among the aquatic plants *Menyanthes*, *Myriophyllum* and *Typha-Sparganium*- type have been observed. Spores of *Polypodiaceae* reach up to 15% and *Sphagnum* – 25%.

According pollen and spores data, sediments accumulated in comparatively cold and wet subarctic – arctic climatic conditions. Local vegetation was very poor. In wetlands, sedge communities with dwarf birch and sphagnum grow. In drier places *Artemisia*, *Chenopodiaceae*, sparse pine – birch forest with some admixture of scattered spruces and juniper grown. The arctic and subarctic plants (*Betula nana*, *Botrychium* and *Selaginella selaginoides*) indicate that, the part of pollen of *Pinus* and *Picea* are transported over long distances. The big amount of pre-Quaternary spores notes that, the pollen grains of *Corylus*, *Quercus*, *Tilia* and *Alnus* are reworked from the older sediments. Low pollen concentration and such plants as *Ephedra* and *Helianthemum* indicate open landscape.

The pollen spectrum from Svirkanciai outcrop similar to spectrum from the interval of Purviai outcrop, which was dated 33800±460 cal BP by ^{14}C. The pollen records show that, the sediments in Svirkanciai site accumulated during Middle Nemunas (Middle Weichselian). So, the sedimentation time could be Mickunai 3 thermomer, tentatively correlated with Denekamp.

ICE-LAKE TYPES DURING THE LAST DEGLACIATION IN NORTHERN FINLAND

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During the Late Weichselian deglaciation Finnish Lapland (northern Finland) was favorable for the formation of ice lakes, since in the main part of the area the variations in level are considerable (100–500 m), the main water divide crosses the area, and the ground level sloped towards the damming glacier. The purpose of this study was to clarify the types of ice lakes, their history and extent, as well as to investigate the conditions at the ice margin. Infra-red color and black-and-white aerial photographs were used in the study. The interpretation was supplemented by field checks, sample drilling and ground penetrating radar soundings.

In the area of high relief, where the glacier was active and its margin unbroken, the ice lakes were small and deep and occurred in valleys between mountains. They are fairly easy to define by the aid of shore marks and spillways. In lower areas the ice lakes were extensive and shallow and filled the present river valleys over areas of thousands of square kilometers. However, the actual size of the lakes cannot be accurately defined, because in many cases it is impossible to determine their extent in the direction of the glacier. Results show that the ice lakes can be divided into five different types. Some of the ice lakes have belonged to two or more types during their history of development.

1. Shallow, marginal and supraglacial lakes.
2. Open ice lake lacking bottom sediments.
3. Open ice lake with little supply of lake sediments.
4. Shallow ice lake with variable thickness of lake sediments.
5. Transition between shallow ice lake and dead-ice area.

Ice lakes of type 1 were shallow basins, formed between the slope of the mountain and the ice edge. Shore marks were infrequently formed because of the small extent and the short duration. Lake sediments are also absent, since these lakes frequently existed partly on top of the ice, as supraglacial lakes. In many cases this ice-lake type was a transition between glaciofluvial stream and pond with stagnant water. The overflow channels between the highest mountaintops are frequently spillways of marginal ice lakes. Mostly they are the only proof of the existence of an ice lake.

Ice lakes of type 2 were water basins filling valleys between mountains, with considerable volumes of open water. They had distinct spillways, which today can be seen as gorges eroded deeply into the bedrock and shore marks such as boulder belts and washed bedrock surfaces reflecting the water levels. No bottom sediments have been found, which is probably due to the strong water current through the lake basin and the absence of a subglacial meltwater conduit, which would have brought sediments to the water basin.

Ice lake type 3 is associated with strong-channeled meltwater action. Today it is distinguished by glaciofluvial landforms on the bottom of the former lake basin. In addition water-transported sand and fine sand were deposited on the bottom of the lakes, and in suitable depressions also thin deposits of fine-grained laminated material. If the ice lake existed a long time or if it received large volumes of glaciofluvial material transported by glacial streams, up to several meters of sediment was deposited on its bottom. In addition to lake sediments distinct spillways and shore marks are also typical of this type.

Ice lakes of type 4 occurred in the river valleys of central Lapland, where they covered extensive areas. Their size varied during their history, as the water level depended on the level of the spillway functioning at the various stages. Several esker sequences cross the ancient lake basins. The meltwater discharging from subglacial conduits laid down sedimentary deposits on the bottom of the basins. The networks of marginal and extramarginal channels leading from one ice lake to the next and the spillways show that the glacier was active and its margin unbroken.

At the end of the deglaciation the margin of the glacier receded to the ice divide zone of Central Lapland. In many places its marginal parts stagnated over an extensive area. In depressions, shallow ice lakes belonging to type 5 were formed. They may have been transitions between shallow ice lake and dead-ice area filled with abundant icebergs or blocks of stagnant ice from the ice margin. Bottom sediments and well-developed shore marks are rare. Instead, the spillways are often very distinct, showing that, however, large amounts of meltwater discharged from the lakes.

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GEOLOGICAL STRUCTURE OF LITHUANIAN QUATERNARY

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Implementation of the constituent part “Spatial geological mapping” of the national geological research program “Geology and Sustainable Development” developed at the Lithuanian Geological Survey, enabled to carry out a genetic and stratigraphic revision of the cross-sections of Quaternary deposits over the territory of Lithuania. The network of reference geological cross-sections (the total number of 45) was developed based on the boreholes with complete stratigraphic sections and identified interglacial deposits (Fig.1). Firstly, twenty-two cross-sections from northwest to southeast (according to the prevailing ice sheet movement direction) were compiled titled I-I’, II-II’ etc. The rest twenty-three cross-sections were compiled from southwest to northeast and titled A-A’, B-B’ etc. These directions of cross-sections were chosen for possibly most comprehensive description of the structure of Lithuanian Quaternary deposits. The sections of 770 cartographic boreholes were used. The cross-sections are spaced 10–12 km. The horizontal scale is 1:200 000 and the vertical scale is 1:2000.

In the subsystem “The Borehole Register”, all boreholes are described using different stratigraphic schemes. Therefore, during the genetic and stratigraphic revision of the Quaternary cross-sections, all stratigraphic units were correlated with the Lithuanian Quaternary stratigraphic scheme (2005).

In the cross-sections the surface's deposits are represented according to R.Guobyte's revised Quaternary geological map at a scale 1:200 000.

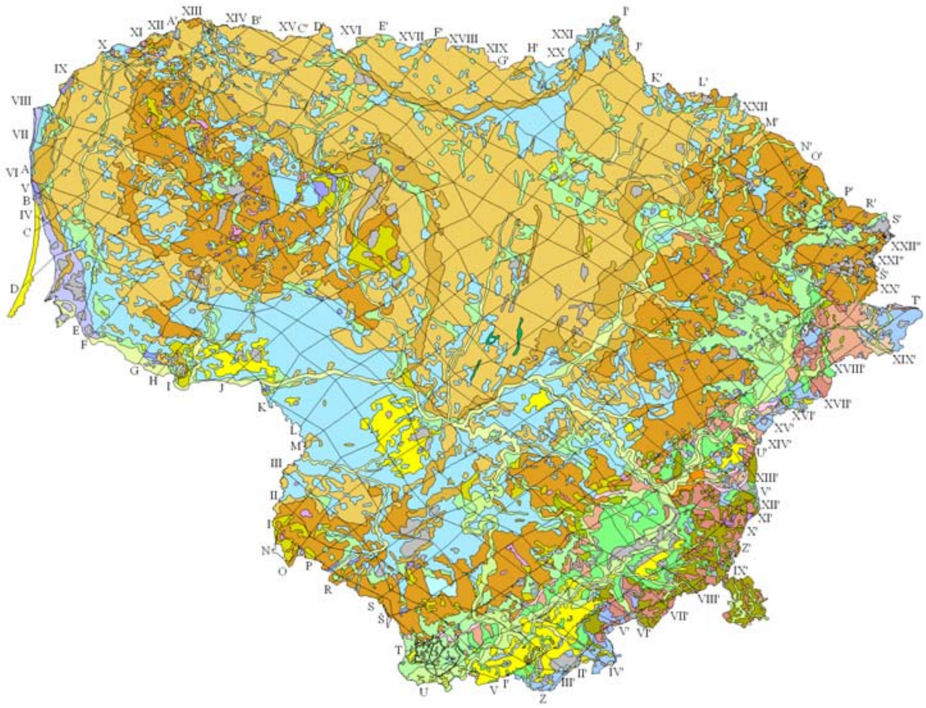

Fig.1. Network of the geological cross-sections in Lithuania.

GROUND-PENETRATING RADAR INVESTIGATION OF PEATLANDS: SELISOO AND RAHIVERE CASE STUDIES

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Recently, two peatlands (Selisoo and Rahivere) have been investigated by Department of Geology, University of Tartu using ground-penetrating radar (GPR). The main objective of the studies was to determine the thickness of peat but also to characterize the topography of sub-peat mineral soil and hydrologic regime (Selisoo bog) and to test and optimize the GPR method for peat resource estimations (Rahivere bog).

The studied cases are different in many aspects including size and type. First, Selisoo is a raised bog in northeastern Estonia with dimensions of 7.4 × 3.7 km. The peat, which mainly consists of poorly decomposed sphagnum peat, attain a maximum thickness of about 7 m. Second, the Rahivere bog is located in Vooremaa drumlin field, eastern Estonia, and is smaller (1.2 × 0.7 km). It developed from a wetland between the drumlins and has been fen during most of the Holocene, and only very recently has become a bog. The peat at Rahivere consist of up to 3–4 m thick well decomposed fen peat that is covered by about 1 m thick sphagnum peat.

Both bogs were investigated with the GPR system Zond 12e by Radar System Inc., Latvia, and using frequency of 300 MHz. The Rahivere peatland was studied during one campaign with a set of parallel profiles 100 m apart whereas the larger Selisoo bog was studied with coarser non-uniform set of profiles. There were four field campaigns into Selisoo trying initially to get a general picture and to refine details at later times. The field works at Rahivere took place in winter 2009; Selisoo campaign included winter and summer seasons in 2007–2009. The GPR profiling was complemented with drillings in both cases.

The GPR records the time that it takes for the emitted electromagnetic (EM) wave to travel to the reflective object and back to the surface. In order to convert time into the depth scale speed of the EM wave (described via permittivity of medium) must be known. The permittivity was estimated (i) by comparing radargrams to drilling data, or (ii) by using common mid-point technique. Both methods were used at Rahivere where permittivity was found in the range 60…70 depending on the extent of decomposition of peat: permittivity of well-decomposed peat was lower than that of poorly decomposed peat. This can be explained with larger amount of water in less consolidated poorly decomposed sphagnum peat. The results from Rahivere are in good agreement with data from Selisoo where comparison of GPR and drilling data yielded permittivity of 72. As mentioned above, the peat at Selisoo consists mainly of poorly decomposed peat.

In radargrams the base of peat can be recognized as reflection with much larger amplitude than the reflections within the peat layer (Figs. 1 and 2). The base of peat can usually be followed over considerable distance. In general, the bog peat has higher inner reflectivity that fen peat, which sometimes looks almost transparent. The reflections within the sphagnum peat in Selisoo are scattered although there are reflective surfaces that can be followed over a distance of hundreds of meters. Based on drillings, such surfaces can be associated with layers of ash, tree stumps and roots or changes in vegetation.

Signal penetration is quite good in both cases. The base of peat becomes discontinues at depth of 6–7 m in the GPR profiles of Selisoo, but the gaps are usually less than 10 m long and the overall picture remains clear. In Rahivere, the contact between underlying lake sediments and covering fen peat can be followed down to 3–4 m in most part of the peatland. However, in the central part, there is about 50 m wide and 1 m deep channel-like feature in which the contact of clayey silt and peat disappears in GPR profiles. We propose that disappearance of signal might be related to higher groundwater salinity within the channel, and in Rahivere in general.
Reflectivity of sub-peat sediments indicating their layering or inner structure is not visible in Selisoo that can mainly by attributed to relatively thick peat layer (typically >5 m). At Rahivere, the silty and sandy sediments on the slopes of depression can sometimes be illuminated within the topmost meter.

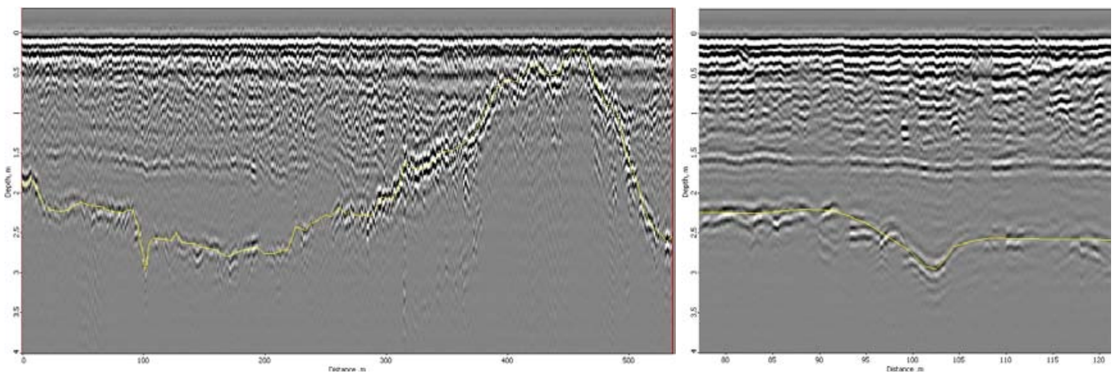

Fig. 1. An example radargram (left) from Selisoo bog showing reflections from the base and from inside the peat layer. The section in the right shows a detailed view of profile in the left.

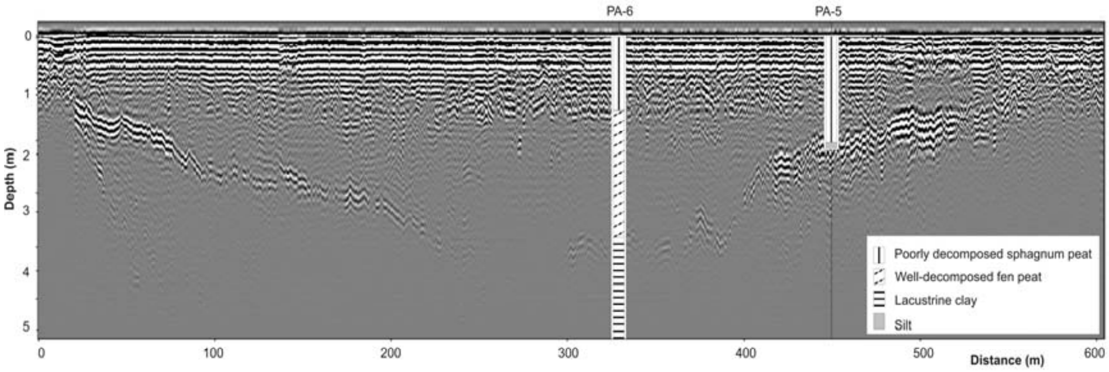

Fig. 2. An example radargram from Rahivere bog with a bowl-shaped reflection that originates from the topmost surface of the mineral soil under the peat body. Please note the vertical scale is exaggerated.

ICE FLOW PATTERN AND EXTENT OF THE LAST SCANDINAVIAN ICE SHEET SOUTHEAST OF THE BALTIC SEA

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Review of scientific literature, geological maps and available previous digital data on ice marginal positions, glacial landforms and sediment distribution are utilised to reconstruct ice streams and lobes, and the maximum (LGM) and subsequent recessional ice marginal positions of the Scandinavian Ice Sheet (SIS) southeast of the Baltic Sea. This paper presents preliminary results of the ongoing research that aims to build a GIS-based model on extent and timing of the last SIS in the area between its maximum extent and the Baltic Sea. Digitized subglacial bedforms, ice marginal and other glacial landforms and features from published sources are compared and validated against digital elevation model (DEM). This has allowed specifying, revising and in places questioning the location of the LGM, and to interpret and correlate the post-LGM ice streams with marginal positions. Morphological evidence persisted from large ice streams in length of 100 to 300 km that operated at the time of formation Middle and North Lithuanian endmorains that is ca 2–3 ka after the LGM. Ice streams from the Onega and White Sea basins had high lateral slopes and clearly channelled flow while the ice streams that drained the ice sheet through the Ladoga – Ilmen depression and in the eastern Baltic had usually fan-shape flow pattern and morphologically unclear lateral slopes. In elevated areas and highlands that are located between ice streams a crossing of lineated bedforms is common. One set of the crossing lineations on Zemaitia, Eastern Kursa, Vidzeme, Haanja, Latgale and Sudoma highlands that are 200–350 km inside of the LGM margin, has north-west – south-east orientation that in general conforms to the direction of ice flow during the LGM and Baltija phases. It is expected that the presented compilation will stimulate discussions and scrutiny of earlier published data and generate ideas for future investigations.

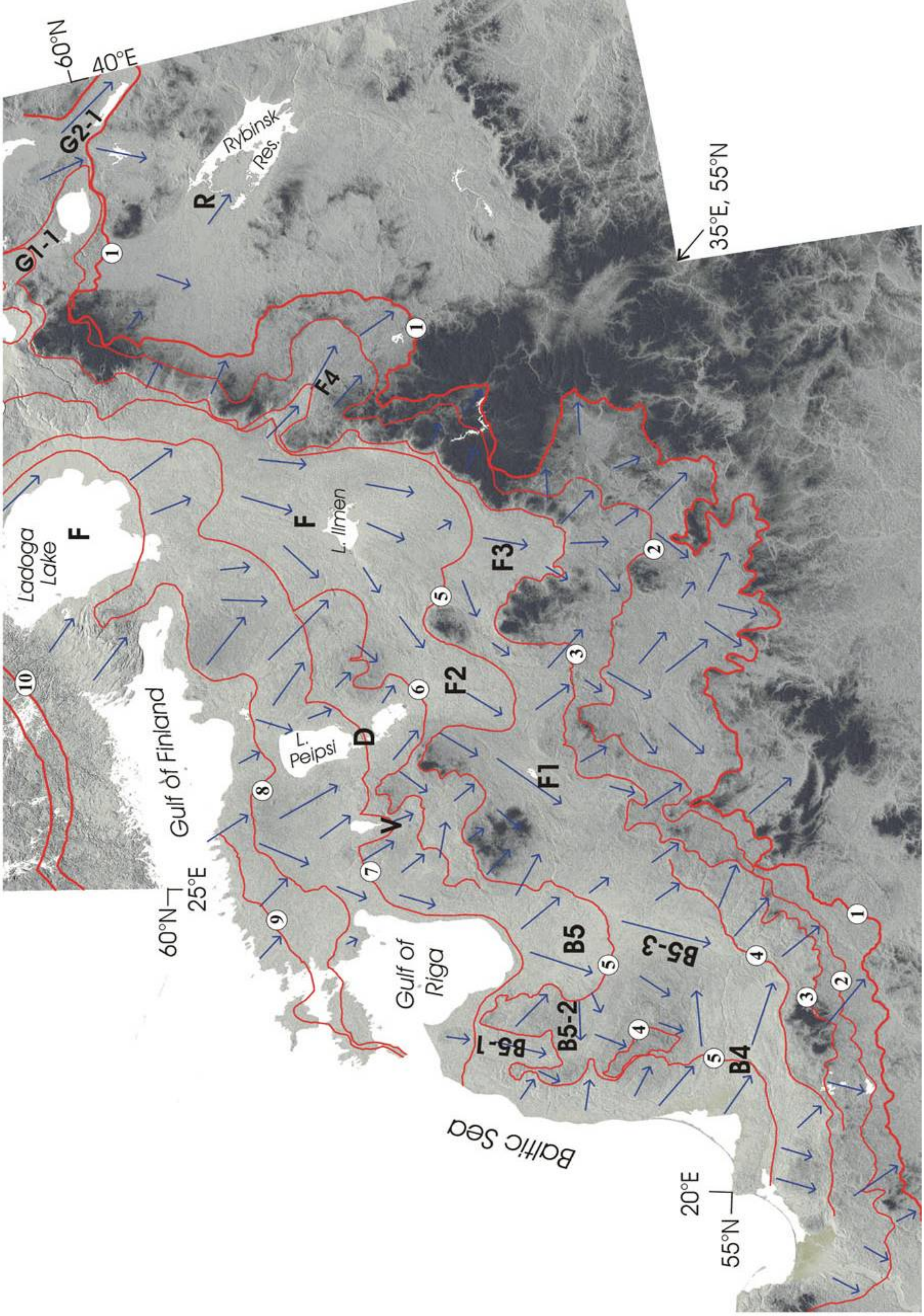

Fig. 1. Ice lobes and marginal positions of the last SIS southeast of the Baltic Sea. Ice streams and their complexes are: Baltic ice stream complex (B) with Neman (B^4) and Riga (B^5) ice-streams and Usma (B^{5-1}), Vadakste (B^{5-2}) and Zemgale (B^{5-3}) sub-ice-streams; Peipsi – Pskov ice stream (D); Võrtsjärv ice stream (V); Karelian ice stream complex with Ladoga – Ilmen – Lovat' ice stream (F) and Lubana (F^1), Velikaja (F^2), Kunja (F^3) and Msta (F^4) sub-ice-streams; White Sea ice stream complex (G) with Beloye Ozero (G^{1-1}) and Kubenskoye (G^{2-1}) sub-ice-streams. Rybinsk (R) ice stream is located outside of the estimated LGM limit.

Ice marginal zones are: 1 – LGM (Gruda in Lithuania), 2 – Baltija (=Pomeranian or Vepsian in Karelia and western Russia), 3 – South Lithuanian (Sebezha and Krestets in Russia and Karelia), 4 – Middle Lithuanian, 5 – North Lithuanian (Haanja and Luuga in Estonia and Russia), 6 – Otepää, 7 – Sakala (Valdemarpils in Latvia), 8 – Pandivere (Neva in Russia and Karelia), 9 – Palivere, 10 – Salpausselkä I (Rugozero in Karelia).

COMPARISON OF THE VEGETATION HISTORY RECORDS FROM THE LAGOONAL LAKE SEDIMENTS IN LATVIA

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Coastal area of Latvia is rich in Littorina lagoonal lakes, particularly central and western coast of the Gulf of Riga, as well as coast of the Kurzeme peninsula. All former Littorina lagoonal areas nowadays are located approximately from 1m to 5–6 m above sea level. Investigation of lagoonal sediments from almost all lagoons indicates their formation during the Atlantic Time. However, there some differences can be found in continuation and interruption of lagoonal sedimentation conditions. Predominantly they can be explained by different altitude of area during the Littorina Sea. In Latvia, in the Littorina Sea sediment section records of two major transgressions (Lit^a and Lit^b) traditionally are recognised. However, there are still several issues to be discussed, e.g. the number of the Littorina Sea transgressions recognised and interpretation of complicated multi-layered Littorina sediment cores in sections from former sea lagoons.

Number of data from ancient coastal lagoons currently occupied by Pape, Liepaja, Bušnieku, Engure, Kaņieris, Sloka and Babīte lakes demonstrate natural changes after the maximum of the Littorina transgression (Lit^a) and illustrates that coastal plain were rich in lagoons, which in the result of sea regression gradually become as coastal lakes. Some of them, like Sārnate and Ģipka, as well as Priedaine currently are completely fill in. Lagoonal sediments in the former Ģipka lagoon are covered by alluvial and eolian deposits, deformed by early farming activities and partly by coastal erosion. Nowadays raised bog is formed in the areas of the former Sārnate lagoon.

Sediments of the Littorina Sea stages in Latvia still are not dated sufficiently and direct correlations are not applicable, therefore still under discussions. Investigations of the Sārnate, Būšnieku Lake, and Priedaine have been carried out during last years, which gave new additional data as well as new ideas and aspects of regularity and similarity of vegetation history recorded by lagoonal sediments.

Plant macro remains and pollen composition of gyttja layer from Ģipka sediments suggest that primary shallow lake had been formed during the first part of the Atlantic Time. Level of water had risen during middle and late Atlantic Time. The grey gyttja in the depth interval 3.8–4.3 m in the borehole No.16a contains seeds of brackish water plants such as *Rupia maritima* and *Zanichellia palustris* characterising the coastal lagoon conditions. The remains of fishes are found in this sediment layer. Algae composition reflects mesotrophical conditions with wide littoral zone during the formation of gyttja. Greenish-brown sandy gyttja is rich in seeds: water plants – 56%, near shore plants – 16%, fen and moist meadow plants – 24%. *Trapa natans* (fragments of endocarps, bristles and pollen) is found. At the same sediment layer has been found also remains of *Potamogeton* and *Najas maritima*. Pollen data also indicate presence of *Trapa natans* as well as, other warm demanding aquatic plant and herb pollen.

The brown gyttja layer in the interval 2.85 – 2.4 m covered by peat has been formed during late Atlantic Time. Upwards gyttja was replaced by fen peat, which point on complete fill-in of lake. The composition of plant remains shows decrease in water plants to 28%, nearshore plants to 10%, while increase of fen and moist meadow pants to 51%. *Alnus glutionosa, Solanum dulcamara, Eleocharis palustris, Menyanthes trifoliata, Cicuta virosa, Carex* spp. and *Lycopus europaeus* are typical.

Complicated sedimentation and vegetation history is characteristic for the Sārnate area. Frequent changes of sea level changes have been recognised. The very lower part at the Sārnate lagoonal area is represented by 20–30 cm thick *Hypnum* peat layer immediately overlies the sandy clay in the large part of present mire area and have pollen spectra (*Betula* maximum) characteristic of the Preboreal, which is confirmed by the date of 8900+–90 (LE-901, Dolukhanov, 1977).

The upper layer of freshwater lime evidently accumulated during the time of the Ancylus Lake in larger part of area, except northern part, where gyttja overlay fen peat. The pollen spectra show the *Pinus* maximum. An ancient shallow lagoon in Sārnate area has been formed because the Littorina Sea transgression, seawater entered the low area between Ventspils and Sārnate through the area of the Venta River mouth. Pollen composition from sediment sequences deposited at that time reflects largest distribution of the broad-leaved trees (pollen zones *Ulmus-Tilia-Alnus*, *Corylus-Ulmus*-Poaceae) in the area characteristic for the Atlantic Time.

During Littorina Sea regression phase the lagoon was cut off from the sea and the gyttja layer formed in shallow lake. This produced pollen spectra with significant proportions of *Quercus* and *Alnus* pollen, and *Carpinus*. Record of *Trapa natans* in the gyttja under the fen peat indicates that during the Atlantic Time *Trapa natans* grew in the lake, which was used for food by Neolithic people.

The shallow lagoon lakes, during the Subboreal time, were separated by low paludified flat lands. Since the second part of the lakes have gradually territorialized and filled-in with fen peat and latter also raised bog vegetation. Raised bog vegetation occurs at that time in the northern part of mire. The Subboreal was characterized by a warm, dry climate, and during this period the Sārnate lagoon lake became territorialized and fen vegetation developed. In the pollen diagram we may trace a significant increase in herb pollen, as well as the *Picea* maximum, indicating the Subboreal Time. During the Subatlantic period, as the climate became cool and wet, the raised bog vegetation developed and *Sphagnum* peat formed in almost entire area of Sārnate Mire.

Pollen diagram from Sārnate Mire reflects vegetation development since the Boreal Time when pine forests were distributed in the area before the climatic maximum. The data obtained and compared with that from earlier studies allow us to find out that raised bog vegetation earlier developed in the northern part of the Sārnate Mire than in the largest par of area, which was occupied by lagoon during Littorina sea transgression phase. Fluctuating pollen curves indicate the impact on the vegetation by the Neolith man. Nowadays, the largest part of the ancient lagoon area is replaced by raised bog although fen vegetation in Sārnate Mire although occurs as well, mainly in the coastal areas of former lagoon.

Pollen and macroremain data from the Priedaine site indicate sediment formation mainly during conditions of shallow lagoon or lake. Fine sand layer on the bottom has been covered by peaty gyttja with wood remains. Sand grain admixture is noticed in gyttja, which amount upwards decrease. Composition of gyttja (amount of sand and plant remains, species of macro remain) frequently changes, what indicate also environmental changes. Pollen data reflect changes in regional and local vegetation composition since the Atlantic Time until nowadays.

New data has been obtained from Būšnieku Lake, which was assumed as lagoonal type lake, which is in endangering by the anthropogenic impact. Results of studies including field works (coring, sediment description), pollen analysis, sediment dating, as well as, analysis of cartographic material, geological map and geological cross-sections of the lake, allow to conclude that organogenic sediments in the Būšnieku Lake have been accumulated since the Preboreal until nowadays without obvious interruptions. The approval for the version of higher level of the Ancilus Lake than Littorina Sea level has been obtained.

Evaluation and comparison the vegetation history records studied from the lagoonal lake (Sarnate, Ģipka, Priedaine, as well as Bušnieku Lake sediments allow to ascertain the similarity of regional vegetation and differences in local vegetation development.

ON THE PROBLEM OF SEDIMENTATION IN THE VENTA PROGLACIAL LAKE

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The glaciolacustrine formations which developed in a dammed glaciolacustrine lagoon occupies about 23% of Lithuania's territory. The glaciolacustrine Venta headstream basin is situated in the western part of Lithuania, at the north-eastern bottom of the divide massif of Žemaičiai Upland. The absolute altitude of the Venta headstream glaciolacustrine basin varies from 100 to 115 m reaching 120–125 m in the peripheral part. The surface of glaciolacustrine plain is undulating and slightly sinking in the south-eastern direction.

Schemes of the thickness of glaciolacustrine sediments and their lithofacies were performed and served as a basis for determining the structure of Venta glaciolacustrine basin, distribution patterns of lithological varieties of sediments and their deposition.

The geological material from boreholes and clay quarries show that the sediments exposed in the Venta headstream basin lie on an uneven till surface. The absolute altitude of the bottom surface usually ranges from 98 to 100 m. However somewhere, especially in the central and southern parts of the basin, the bottom sinks to 87–96 a a. The lithological composition and thickness of glaciolacustrine sediments depend on the depth of glaciolacustrine basin and deposition area. The thickness of sediment layer mainly has been predetermined by the irregularities of the bottom of periglacial lake and the amount of transported material. The thickness of sediment layer in the basin is rather uneven. The dominant thickness is 5–10 m. The bottom of the Venta headstream glaciolacustrine basin is furrowed by deep trough-shaped old valleys and depressions used by Venta, Knituoja, Aunuva and Ubesiukas rivers.

The Aunuva valley is trough-shaped and reaches 0.8–1.5 m. width. It's upper part is composed of sandy silt clay with scarce gravel somewhere merging into silt or fine-grained sand with gravel. Silt and various grained sand indicate a shoaling basin. The thickness of these sediments ranges from 3 to 8 m. They overlie brown, greyish brown, dark brown and greasy varved clay. Alternating clay and silt or silt clay layers represent the varves. Clay layers are dominant. The thickness of silt layers varies from 0.5 to 2 cm and the thickness of clay layers from 1–2 to 5 cm. The content of clay particles in the clay is rather high: 60–70 % or somewhere even 85 %. The content of clay particles is even slightly higher in the lower part of the layer. The thickness of clay layer is uneven and ranges from 1.7 to 14.0 m. It tends to increase in the direction of the confluence of Aunuva and Venta rivers where it reaches 21.7 m.

A sketch map of sediment lithological groups in the Venta headstream glaciolacustrine basin has been compiled based on the available borehole data. It demonstrates the lithological composition of dominant glaciolacustrine sediments in every point of the basin (Fig. 1).

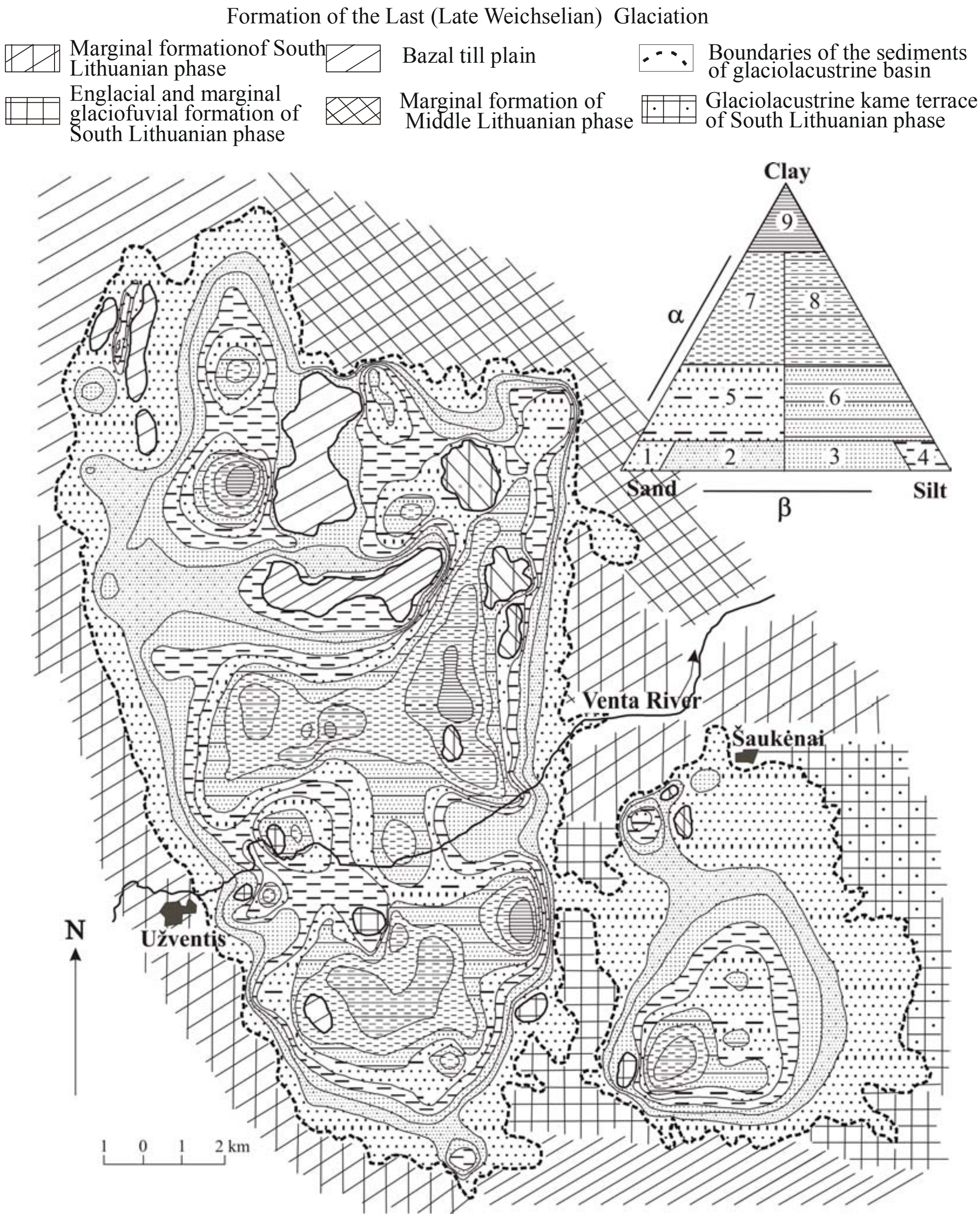

Fig. 1. Distribution scheme of lithological groups in the Venta River headstream glaciolacustrine basin. Description of lithological groups (1–9) and conventional signs are given in Table 1.

Table 1. Explication of the triangular diagram of lithological groups

Field No	Lithological composition (groups)	Clay coefficient α	Sand coefficient β
1	Sand	<0.25	>8
2	Silty sand (sand dominate)	<0.25	8–1
3	Sandy silt (silt dominate)	<0.25	1–0.25
4	Silt	<0.25	<0.125
5	Clayey sand (sand dominate)	<0.25–1	≥1
6	Clayey silt (silt dominate)	<0.25–1	<1
7	Sandy clay (clay dominate)	1–8	≥1
8	Silty clay (clay dominate)	1–8	<1
9	Clay	>8	Anything number

GEOLOGICAL AND GEOMORPHOLOGICAL STRUCTURE OF THE GLACIOLACUSTRINE KAME TERRACE OF THE NORTH LITHUANIA PHASIAL, BALTIJA STAGE

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Glaciolacustrine kame terrace are found near the distal slope of peripheral deposits the North Lithuanian phase in the Judiškiai – Megučionys – Linkuva – Vienžindžiai environs (Fig. 1). The 22 km long and 0.5–1.0 km width glaciolacustrine kame terrace on the distal part of Linkuva ridge was mapped and investigated in detail. Its surface, slightly inclined southward, reaches 65–70 m in absolute height. The northern edge of the terrace is bound by the marginal ridge, which rises more than 80 m above sea level.

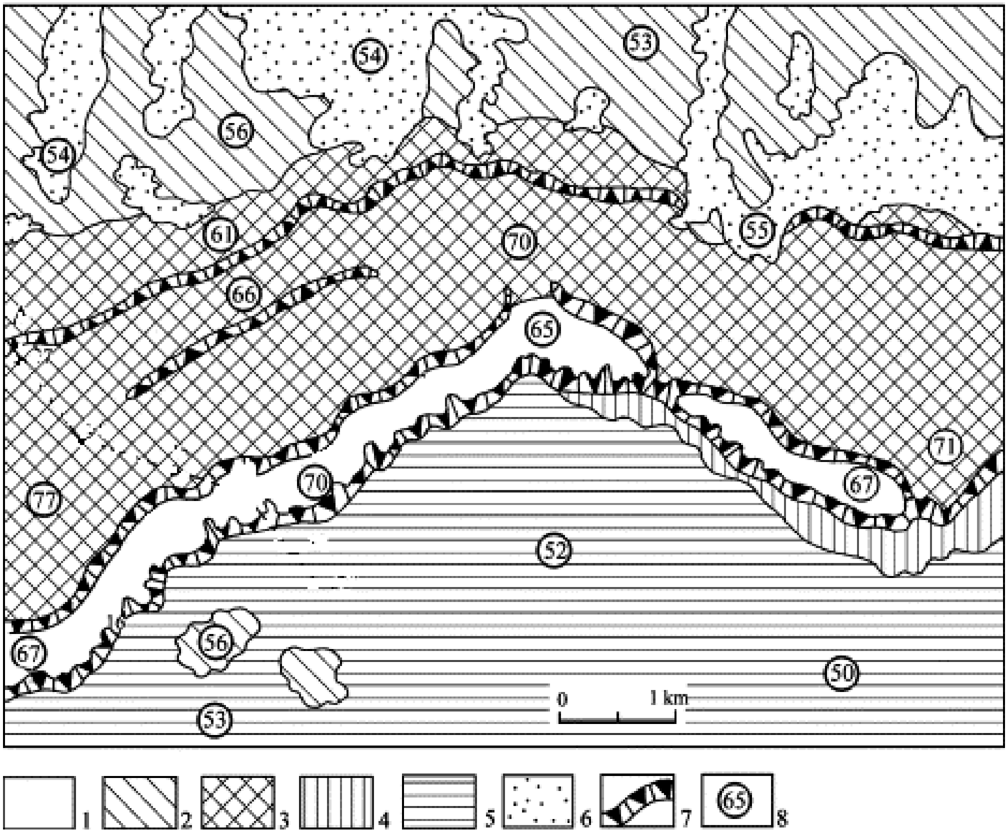

Fig. 1. Geomorphologic scheme of marginal morainic ridge and glaciolacustrine kame terrace, east from Linkuva. (1) Glaciolacustrine kame terrace; (2) basal till plain; (3) marginal morainic ridge; (4) solifluction sheet; (5) glaciolacustrine plain of proglacial lake; (6) plain formed by sediments of different genesis (glaciofluvial, glaciolacustrine, lacustrine, etc.); (7) steep slope; (8) prevailing altitude of relief, in metres

The surface of the kame terrace is 10–15 m above the glaciolacustrine plain of ice-dammed lake (sediments presented generally by varved clay) that is located to the south of the terrace (Fig. 2.).

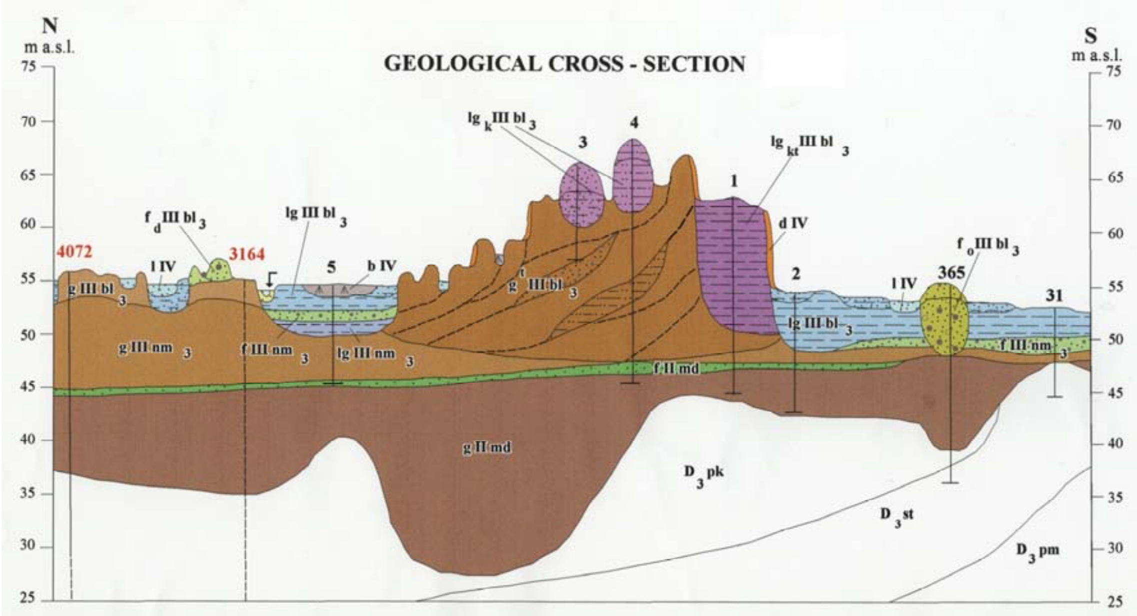

Fig. 2. Geological cross-sections via marginal morainic ridge and glaciolacustrine kame terrace close to Linkuva.

Ravines intersect the slope of the kame terrace. The solifluction debris is developed in the eastern part of terrace foot. In general, the kame terrace consists of glaciolacustrine sediments; its fixed maximal thickness is 15.3 m. Glaciolacustrine sediments are presented by brown clay of massif structure (or silty clay), in some places with small lenses (up to 1–2 cm thick) of fine-grained sand. Clay particles (fraction < 0.005mm) make 70–75%. Silt particles make only 20%. The rest, sandy fraction make 10%. Gravel and carbonaceous concretions (diameter up to 10 mm) in clay sediments is up to 3–5% (Fig. 3.).

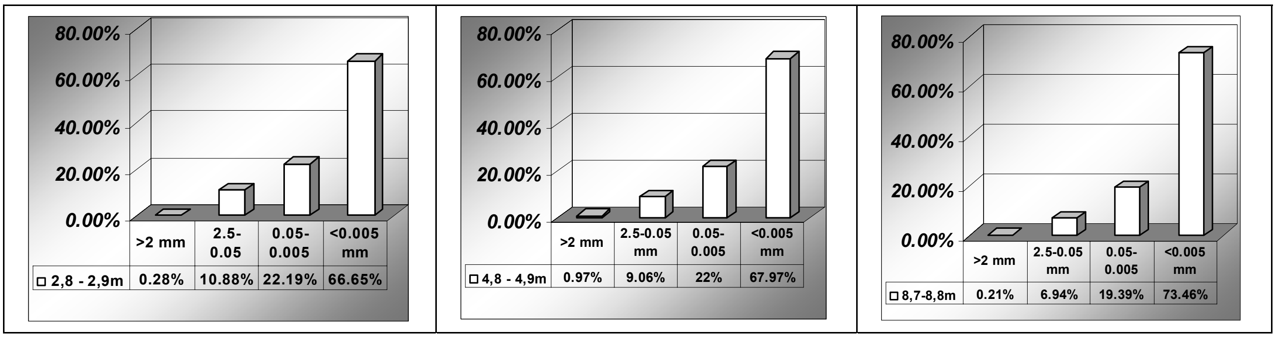

Fig. 3. Granular compositions of kame terrace sediments in borehole No 44.

Kame terrace deposits were developed between the glacier and a neighboring valley slope. Geological structure of Linkuva ridge indicate that this ridge and adjoined kame terrace were formed between active ice lobe and field of died ice dislocated at front.

LANDSLIDES IN PROGLACIAL VARVED CLAYS OF WESTERN ESTONIA

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During the last decade the frequency of landslides at river valley slopes eroded into the glaciolacustrine plain in western Estonia has grown considerably. We studied in detail 9 recent landslides out of 25 known and recorded sliding events in the area. All landslides occurred at the river banks in otherwise almost entirely flat areas of proglacial deposits capped with marine sands. Glaciolacustrine varved clay is the weakest soil type in the area and holds the largest landslides. According to the slope stability modelling and field surveys the critical slope angle for the clay slopes is ≥10º and for the marine sand slopes ≥20º. Fluvial erosion is the main process in decreasing slope stability at the outer bends of the river meanders. An extra shear stress generated by groundwater flow or rapid water level drawdown in the river channels are responsible for triggering the small scale landslides right at the river bank. Those small scale slides may lead to larger landslides upslope, therefore forming retrogressive complexes of slides in the glaciolacustrine clay. The biggest investigated landslide, moving in total 10 100 m^2 of the slope, was a complex of at least six sliding events and occurred in 19.12.2005. Regular topographical surveys show rapid erosion of the formed landslide toe and appearance and growth of the new landslides in already ruptured slope. Slides in marine sand are single events and the triggering factor is the additional shear stress, generated by water seepage that destabilizes the upper portions of slopes. Therefore those events occur always after the heavy rains or thaw. Landslide susceptibility map, covering the area around Pärnu town, was composed based on digital elevation model (4x4 m cell size), the data on spatial distribution of glaciolacustrine clays and marine sands, and on existing and critical slope angles of these deposits. According to this map the Sauga river valley is the most prone to the slope failures.

CARBONATE PRECIPITATES WITHIN GLACIOFLUVIAL DEPOSITS IN LATVIA

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This study examines the sedimentary and morphological settings of secondary carbonate precipitates within glaciofluvial complexes in Latvia. The carbonate precipitates were found in five outcrops in the central or peripherial zone of the accumulative insular uplands (Fig. 1). Calcite cement occurs mostly in glaciotectonically deformed glaciofluvial complexes overlain discordantly by till. In all studied exposures cemented masses occur above the modern water table.

Calcite cement distribution and the level of the cementation is not uniform. The cemented masses are not laterally continuous for more than a few meters. Carbonate precipitation has been taken place close to the contact zone between till and glaciofluvial deposits intruded also into the till for some tens centimeters. According to Āboltiņš (1989) the formation of the calcite cement probably occurred during or shortly after the glaciotectonic deformation due to migration and discharge of the subglacial pore-water along the thrust sheets as suggested by more cemented layers at the thrust sheet soles.

Thicker cement crusts (up to few mm) are commonly visible in coarser, matrix-free facies of gravel. In matrix-rich sandy facies the cement is distributed in the matrix filling almost overall intergranular pore space. Micro-morphologically this matrix-supported cement is present mostly as micrite (<4 μm) and microsparite (4–10 μm) angular equant to rhombohedron calcite crystals. Micrite is usually located next to grain boundaries indicating that micrite preceded microsparite or sparite precipitation. Sparitic (up to 50 μm) calcite assemblages fills larger intergranular space or in microfractures. Multilayered cement and dissolution traces on the surface of calcite crystals suggest episodes of cement precipitation alternating with periods of dissolution. These cements are interpreted as having formed in water-saturated conditions based on the continuous and isopachous character.

The study will further expand on the palaeoenvironmental interpretation derived from the isotopic composition and the age of carbonate precipitation in the Peribaltic area. The study is funded by the Estonian Science Foundation research grant No 6962 and the target-financed project SF0182530s03.

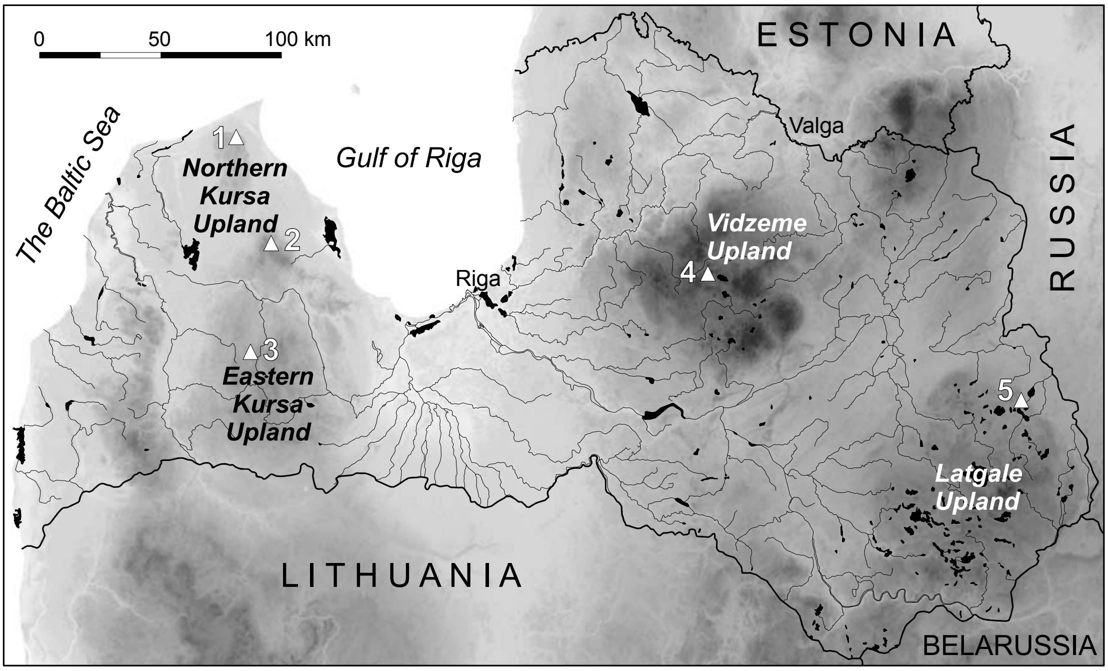

Fig. 1. Location of studied sites: 1 – Vidāle (esker), 2 – Lejaslabiņi (composite hill), Satiķi (bedrock dolomite raft), 4 – Laidzu (plateau-like hill), 5 – Čodorāni (esker).

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LATE GLACIAL TO HOLOCENE LAKE BASIN AND DRAINAGE PATTERN DEVELOPMENT WITHIN POMERANIAN ICE MARGINAL ZONE (NE-GERMANY)

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The presented investigations deal with the Late Quaternary development of connected lakes and rivers within the Pomeranian ice marginal zone (W2), the Early Pomeranian advance (W2max) and the adjacent outwash plains in central Mecklenburg (NE–Germany). The investigations achieve a combination of geomorphical, pedological, palaeobotanical, palaeolimnological, archeological and historical results according the palaeohydrographic reconstruction of lake–river systems since the Weichselian Late Glacial. Multi proxy studies of lacustrine sediments in connection with pedological and geomorphological investigations of surrounding lake fringes are still a very unique topic in regional surface water studies and thus they offer new results and supplement older ones. The studies emphasises the relationship between lake basin and river valley development, the sedimentation history within the lake basins, the lake-level development and the interaction between man and lake. The field works focussed on the Lake Krakower See (47,5 m a.s.l., 16,5 km²), the Lake Woseriner See (37,1 m a.s.l., 2,7 km²) and the Lake Drewitzer See (≈61 m a.s.l., 6,92 km²) as well as on near catchment areas like lake terraces. Furthermore the incised valley of River Nebel (W2, adjacent to Lake Krakower See) and the incised valley of River Mildenitz (W2max, adjacent to palaeolake) were investigated by field works (Kaiser et al. 2007, Lorenz 2007).

Lake terraces and fossil lacustrine sediments surrounding several lakes give evidence for highest lake-levels during the Late glacial. The Lake Krakower See has existed at the latest since Bølling (12.700–12.150 BP) as a more shallow and as a twice as big palaeolake with a lake-level of 51 m a.s.l. Palaeoenvironmental studies on basal deep water lacustrine sediments gave rise to initial shallow water conditions for Lake Woseriner See and Lake Drewitzer See despite the high lake-levels. The lacustrine sedimentation started here during Allerød (11.800–11.000 BP) with strong organic silicte gyttja in presence of buried dead ice. Actually both lakes reach more than 30 m of water depth. The Late glacial–Early Holocene transition is documented within the facies of lake basins and basin fringes very well. This transitional phase induces the most remarkable relationship between the incised valleys penetrating the ice marginal zones and the adjacent lake basins, because the north headed drainage system established not before Younger Dryas triggered by final dead ice melting. While the incised valley of the River Nebel has no older river terraces and influenced the Lake Krakower See, the incised valley of the River Mildenitz has five river terraces and influenced Palaeolake Dobbiner Plage. Both valleys initially descended by subglacial channels. The Early Holocene drainage caused an incision of the valleys, the connection of small and seperated lake basins formerly sealed with dead ice, the sedimentation of thick fan deltas within the now connected lake basins and a remarkable lake-level lowering in lake basins situated south to marginal zone. Preboreal basal peats at lake fringes of the Lake Krakower See and the Lake Woseriner See are sedimentological evidence for Early Holocene lake-levels 3–5 m below the recent ones. The bigger lake basins also reach maximum basin volumes because the last dead ice vanished during the Preboreal (10.000–9.000 BP). From the beginning of the Boreal (9.000–8000 BP) the lake-levels had been rising continuously until the Younger Atlantic (around 6.000 BP) – in Lake Krakower See by c. 3 m in a timespan of 3500 years. Mid-Holocene peat aggradation at lake fringes also started with the

Younger Atlantic in Lake Drewitzer See and a little later in Older Subboreal (around 5.000 BP) in the Lake Krakower See. Synchronously the first neolithic settlement indicators occur in the pollen diagrams. Between 4.000–2.800 BP (Middle–Younger Subboreal) a lake-level lowering is proofed for Lake Drewitzer See by sedimentological, archeological und palaeolimnological means covering the middle Bronce age. The transition to the Subatlantic around 2.500 BP is characterised by a new lake-level rise and the accumulation of thick beach ridges at the Lake Krakower See and the long-range drowning of peat fringes at Lake Drewitzer See. During the Middle Subatlantic (after 1.300 BP) a clear lake-level low stand occurs for about 300 years reflecting the medieval climate optimum. Hence the 13th century AD the general trend of repeated rising of lake-levels is superimposed by the ejects of medieval watermill stowage. A switch to higher trophic status is connected with intensified medieval land use, clearing of woods and watermill stowage. Furthermore the lake sediments show a distinct increment of silicate components since the 13th century AD. Between 13th to 18th century AD the lakes show a higher lake level than today, but they are also subject to strong seasonal variations and strong influence by watermills. Since the 18th century AD surface waters have been gone through a considerable transfiguring by melioration, land improvement, canalisation, and anthropogenic lowering.

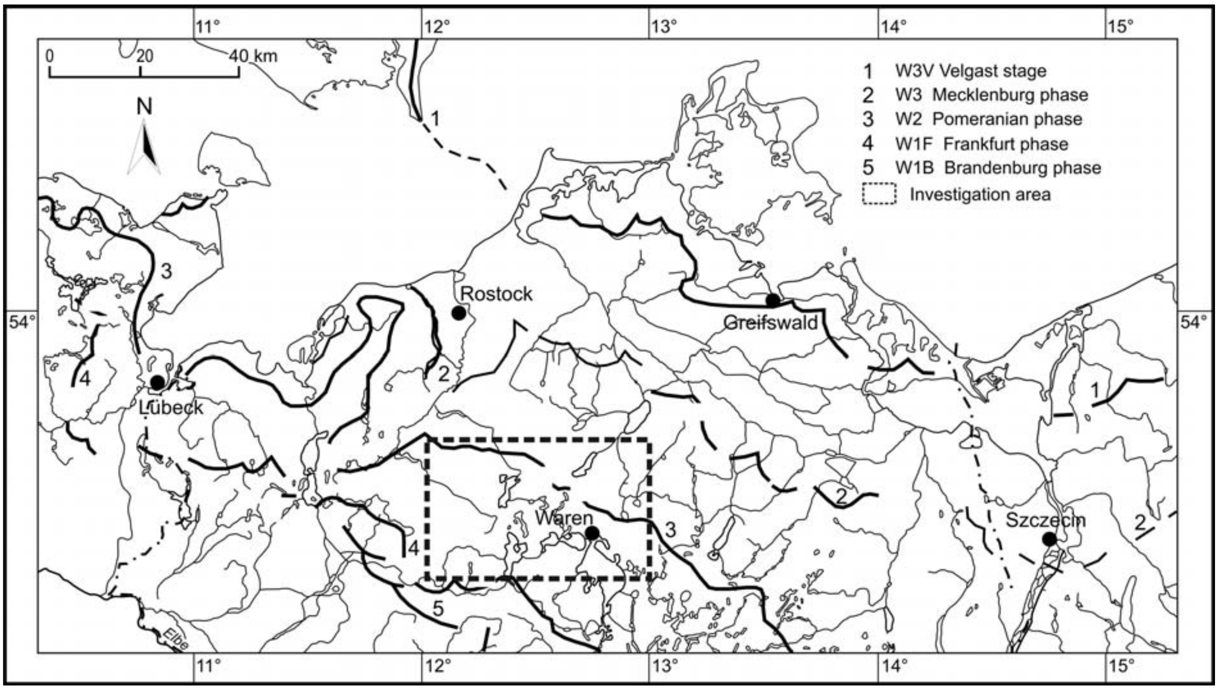

Fig. 1. Ice-marginal zones in NE-Germany and location of the investigation area.

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THE LAST GLACIAL HISTORY OF NORTH-EASTERN GERMANY – FROM MORPHOSTRATIGRAPHY TO GEOCHRONOLOGY

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During a research project in the young moraine area of North-Eastern Germany, samples of fluvioglacial sands have been taken from outwash plains (sandar) with a clear connection to distinct ice margins. Optically Stimulated Luminescence (OSL) dating techniques, primarily based on the single aliquot regeneration dose protocol (SAR) for the dating of quartz, have been applied in order to work out an absolute chronology for the glacial sequence.

The classification of the Weichselian Pleniglacial in North-Eastern Germany has mainly been based on the morphostratigraphical interpretations over the last 130 years. In some places, new results from this OSL dating project do not agree with the established pattern. The unambiguous geomorphological differentiation of the main ice marginal positions (Brandenburg, Frankfurt, and Pomeranian Stage) and their geochronological position turned out to be somewhat questionable.

The relatively weakly developed ice marginal position of the southernmost ice advance of the Brandenburg Stage has traditionally been ascribed to the LGM. Ages obtained from OSL measurements indicate that this fast paced and short lived ice advance might have occurred in the early OIS 2 and conformed to the morphology from the penultimate (Saalian) glaciation. The Frankfurt stage, a long-lasting marginal position during the down wasting of the OIS 2 glacier, seems to be a patchwork of landforms of different ages rather than a contemporaneous formation. The most prominent ice marginal position in North-Eastern Germany is that of the Pomeranian stage. OSL dating results suggest that the corresponding ice advance occurred at around 23–20 ka BP.

A critical review of the morphostratigraphically based chronology and its history against the background of these new results from OSL dating will be presented at the meeting.

PALYNOLOGICAL CHARACTERISTICS OF DEPOSITS OF THE BYZOVAYA (MIDDLE VALDAI) HORIZON IN THE CHORNAYA RIVER BASIN, NORTH OF THE EUROPEAN PART OF RUSSIA

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The Byzovaya (Middle Valdai) horizon is widely distributed in the Chornaya River basin and it is exposed in the coastal outcrops. The horizon overlies the Moscow till, the polar till or Holocene deposits being superposed. Limnic, lake-bog and alluvial sediments with the interlayers of peat represent the horizon. The age of the Byzovaya horizon deposits is established by the radiocarbon datings within 48–25 thousand years, the fossils of mammal's fauna and by typical palynological complexes.

The study of some sections (Andreicheva & Duryagina 2005, Andreicheva, Marchenko-Vagapova & Golubeva 2008) have allowed to receive the picture of changes of the palynological complexes and to allocate following phases of the vegetation changes, replacing each other:

The phase of vegetation Bz_I. During this period, the open forests mostly by birch where pine and fir occupy the subordinated position existed. The scrub and the grassy associations formed by *Chenopodiaceae, Poaceae*, *Artemisia* sp., *and Ephedra* sp. have been developed. The distribution of the bog-tundra associations and the presence of xerophytic species characterise a dry and cold climate.

The phase of vegetation Bz_{II}. At that time, the thinned wood groupings were distributed. In the basin of the Chyornaya River, the birch is dominant. Fir and a pine were in considerable quantities in those complexes. Together with the boreal (*Ericaceae*, wood *Lycopodiaceae* and *Polypodiaceae*) flora elements, the bog and meadow (grasses, species of genus *Brassicaceae, Ranunculaceae, Caryophyllaceae*) formations the xerophytic species also participated in insignificant quantities. The climate was humid and boreal.

The phase of vegetation Bz_{III}. Palynological spectra of this zone testify that boreal wood communities have lost their dominating position. The light forests formed by birch, pine, fir, and alder have substituted them. The bog-tundra formations along with the xerophytic communities formed by *Artemisia* sp., *Chenopodiaceae, Poaceae* were widely distributed. The climate was cold enough and humid.

The phase of vegetation Bz_{IV}. The various thinned wood groupings: birch, birch-fir, birch-pine were the basic component of the vegetative cover. The meadow vegetation occupied open spaces, which is reflected in the content of pollen of grassy plants (pollen of grass and herbage: *Polygonaceae, Brassicaceae, Ranunculaceae, Rosaceae, Asteraceae*). The climate was cold, but warmer, than in the previous phase.

The phase of vegetation Bz_V characterises the development of birch light forests. Underbrush and grassy associations of the open places composed by *Artemisia* sp., *Poaceae* and herbage have become widely distributed. Further, the bog-tundra formations along with the xerophytic communities have grown considerably. Probably, some cooling of the climate occurred.

The phase of vegetation Bz_{VI}. Spore-pollen spectra of that time point to some climatic warming and to the formation of the thinned birch-fir, birch-pine wood communities. The meadow complex formed by grasses and mesophilic herbage (species of *Polygonaceae, Onagraceae, Ranunculaceae, Rosaceae, Caryophyllaceae, Cichoriaceae,* and *Asteraceae)* occupied the open spaces.

The phase of vegetation Bz_{VII} characterises a new cooling of the climate. There were widespread communities scrub tundra and small open places of birch and fir woods. The absence of the representatives of the xerophytic forms, the participation of grasses and mesophilic herbage (*Ranunculaceae, Brassicaceae, Rosaceae, Polygonaceae, Caryophylaceae* and *Asteraseae*) testify to the development of the tundra communities. The complex characterises the conditions similar to the arctic tundra.

The given phases of the vegetation specify the absence of the climatic optimum in the Middle Valdai (Byzovaya), but, nevertheless, it is possible to identify three periods of warming (Bz_{II}, Bz_{IV}, Bz_{VI} vegetation phases). They reflect cooler climate and characterise the distribution of the thinned wood groupings, the open spaces of which were occupied by marsh and meadow formations.

This research was supported by the grant from the Division for Earth sciences of RAS (№ 12).

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PHASE MODELS AND CONCEPTS OF SAALIAN ICE COVER IN THE NETHERLANDS AND NW GERMANY

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During MIS6 in the Saalian The Netherlands and Germany were partly covered with an ice sheet. This ice cover caused a very distinctive geomorphology with large thrust moraines, sandur plains and glacial basins. In the last 50 years several phase models of glaciation in both the Netherlands and Germany were developed. Some of these models will be discussed during the presentation. Most of these models are outdated now; the last model was made in the early nineties.

My MSc-research focuses on describing the Saalian glacial morphology that is present in both The Netherlands and Germany and combining this in an interactive GIS-model. This is done to get a proper overview of the glacial features that are present and to link several Dutch and German researches. A second aim is to develop a new glaciation phase model combining the geomorphological description and new insights from glaciology. A concept version for this new glaciation model will be presented.

MIDDLE WEICHSELIAN INTERSTADIAL DEPOSIT IN THE VESKONIEMI, NORTHERN FINLAND

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Geological Survey of Finland (GTK) is executing a project, which aim is to update the Quaternary chronology of northern Finland during 2008–2011. New data is mainly provided by studying the stratigraphy and sedimentology of the test pit sections and by using Optical Stimulated Luminescence (OSL) dating method for the inter-till stratified layers (Sarala 2008). In this abstract the results of one test site are described. It is located near the Lake Inarijärvi in the place called Veskoniemi in central Lapland (Fig. 1).

In the sampling site 1 (Fig. 2) a gray till deposit with fissility structures is found in the top of the stratigraphy (0–1.8 m). Under that, there is a massive till layer to the depth of 2.3 m. Laminated sands with some cross-beddings are followed in the stratigraphy from the depth 2.3 m to 3 m, chancing to the deformed sand deposit at the bottom of the section (3–5 m). Totally four OSL samples were taken from the upper part of the sand deposit, two parallel samples from the depth of 2.4 m and 3 m.

In the sampling site 2, the stratigraphy begins with a gravel deposit in the surface (0–0.6 m). Under that there is a layer of medium silt with small stones to the depth of 3.5 m, containing following interlayers: laminated fine sand (1–1.1 m), laminated/rippled sand (1.15–1.4 m and 1.6–1.75 m). Samples for the dating purpose were not taken from the sampling site 2.

OSL samples were dated in the Luminescence Research Laboratory of the Radiation Research Department at Risø National Laboratory, Denmark. Ages of the sand deposits based on the OSL dating were 21 ± 1 and 22 ± 2 ka for the laminated sand layer in the depth of 2.4 m. At the depth of 3 m the ages of the sand layers were 46 ± 3 and 39 ± 3 ka.

GTK has done some earlier stratigraphical studies in the area (e.g. Hirvas 1991). According to till fabric studies the latest ice-flow direction is from NNW to SSE (Fig. 1). The till fabric in the Veskoniemi is not typical for the area, as the latest ice-flow direction has been mainly from SW to NE. Exceptional ice-flow directions are probably due to the separation of ice margin into the smaller ice-lobes with variable flow directions during the end of the deglaciation. Anyway, the dating results are telling that the till covered sand layer is deposited during the Middle Weichselian, even if the dating results from the upper sampling level are somewhat younger. The Middle Weichselian interstadial deposits from the central Lapland have also been described by Helmens et al. (2007), giving OSL date of 48 ± 16 ka for the sediment

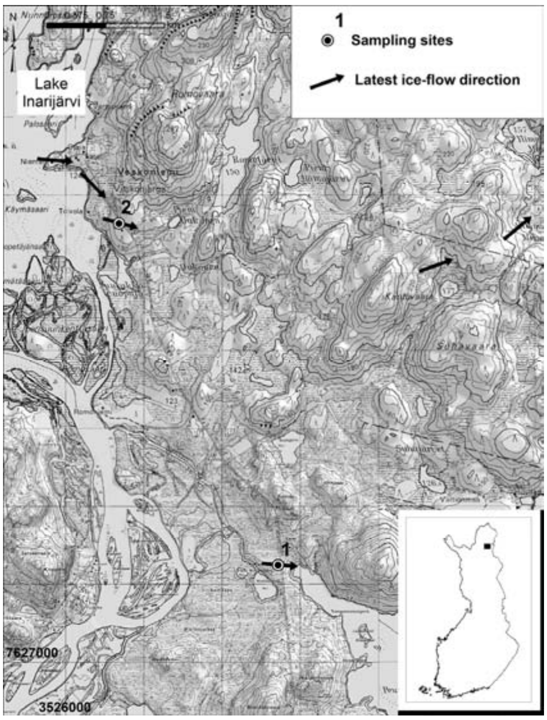

Fig. 1. Map of the sampling sites and the latest ice-flow directions after Hirvas (1991). Base map © National Land Survey of Finland, permission No. 13/MYY/09.

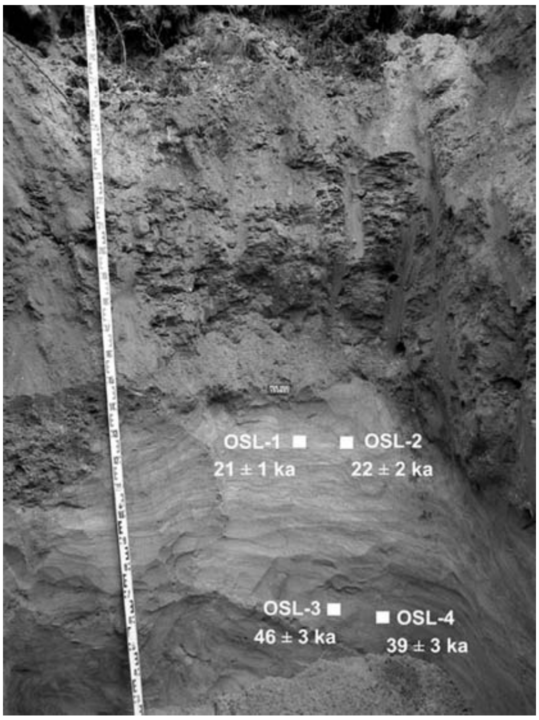

Fig. 2. Sampling site 1 with OSL dating results.

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DEVELOPMENT OF PALAEOINCISIONS OF SUB-QUATERNARY SURFACE DURING THE PLEISTOCENE IN SOUTH LITHUANIA

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Interrelation between the sub-Quaternary surface and the modern one remains an interesting and relevant question. According to the new data, the direct link between these two surfaces was determined.

Analysis of the sub-Quaternary surface based the conclusion that indicated scarps, slopes and palaeoincisions are of different age and genesis. The formation of scarps and main palaeoincisions has been preconditioned by the neotectonic structures mainly.

Valleys of ancient palaeo-rivers later affected and transformed by the glacier melt water and erosion processes was a background for the further development of the sub-Quaternary palaeoincisions. The new elements of various genesis might have appeared in this network throughout the Quaternary as well (Gaigalas et al. 1976). Formation of the deep segments of the palaeoincisions should be associated with the mentioned processes. Normally sub-Quaternary palaeoincisions reach down to 30–80 m in depth and may be accepted as usual in comparison with the depth of the present river valleys (Šliaupa 2004).

Collected valuable data describing the lithology of material filling up the palaeoincisions and covering watersheds, increasing precision of the hypsometric picture of the sub-Quaternary relief, together with valuable experience obtained during the former investigations carried out in the region (Baltrūnas, Pukelytė, 1998) based the compilation of a new more detailed palaeogeomorphological scheme of Lithuania territory. The palaeogeomorphological regions with particular lithomorphogenetic characteristics have been determined in the sub-Quaternary surface of the Lithuania territory. There are four structural-palaeogeomorphological regions indicated in South Lithuania: Great Lithuanian Lowland (II), South Lithuania Plain (IX), Šventoji Slope (VIII), East Aukštaičiai Plateau (VII) (Fig. 1).

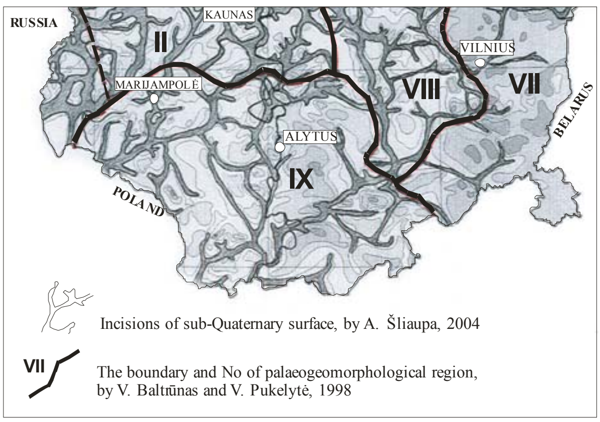

Fig. 1. Palaeoincisions and palaeogeomorphological regions of South Lithuania.

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ON THE AGE OF ESTONIAN PLEISTOCENE DEPOSITS

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About 95% of the Quaternary cover in Estonia is formed of glacial and aqueoglacial deposits and their age is unclear. Five till formations, often of a considerable thickness, are traceable (Raukas 1978). Only in a few cases, they are separated from each other by deposits containing spores and pollen of interglacial or interstadial origin, which considerably aggravates the correlation and dating of glacial strata. Continental stratotype layers of Eemian (Rõngu) and Holsteinian (Karuküla) interglacials are redeposited, what complicates the problem (Raukas & Kajak 1995).

Unclear is also the deglaciation history. The territory of Estonia was freed from the continental ice in Gotiglacial time, which was characterised by high activity of the glacial streams and lobes and a highly variable glacial morphogenesis (Raukas et al. 2004). In the course of deglaciation, active, passive and dead ice conditions probably underwent a gradual change depending upon the climate and bedrock topography. Flow tills and collapses of ice cavities were common in dead-ice areas, which is why organic layers with different ages often occur in an unusual stratigraphic position.

In the study area, traditionally five ice-marginal belts have been distinguished: Haanja, Otepää, Sakala, Pandivere and Palivere. In the light of pollen analytical interpretations, the retreat of the ice margin from the Haanja zone (the oldest in Estonia) began in the Bølling, whereas Estonia was finally free of ice in the second half of the Allerød chron (Pirrus & Raukas 1996).

Deposits of the huge ice-dammed lakes give some hope to use varve chronology with the simultaneous study of secular variations of the geomagnetic field. However, in reality, annual layers are far from being always identifiable, and in many cases the seasonal lamination is frequently accompanied by the arrangement of material in layers within one annual cycle. In addition, as proglacial lakes were isolated from each other and many gaps exist between the neighbouring territories, the accuracy of the estimated rate for the ice recession in the areas where glaciolacustrine sediments are absent is extremely disputable. Probably the retreat of the ice alternated many times with short intervening new advances. It would be very difficult or even impossible to estimate how many years it took for the ice margin to retreat from a chronologically dated line to the north and readvance back to a new line. It is difficult to correlate Estonian varve countings to the Swedish varve chronology, which is also changing from time to time. In varve years Estonia became ice free at about 14 000 – 12 600 years ago (Kalm 2006).

Application of the radiocarbon method is also limited due to the absence of good interstadial sections for dating, redeposition of organic matter and contamination with young and ancient carbon. The majority of the radiocarbon dates obtained from submorainic and intermorainic sequences are younger than one would expect on the basis of the conventional radiocarbon method. The application of the ESR method is limited due to the lack of good dating material.

In last decades, widely TL and OSL methods have been used for the dating of aqueoglacial deposits, but at least for Estonia the obtained data are extremely heterogeneous and most of them are entirely unreliable. From one side inconsistent results are depending from geological factors, because the deposits have accumulated in different sedimentological conditions and often redeposited, but with no doubts physical grounds of study methods need significant improvement also.

Application of the ^{10}Be method to establishing boulder exposure ages on the top of end moraine belts yielded highly variable results, because it is not known how long the investigated boulders have been in the forest, under snow cover or below the waters of the Baltic Sea. Nevertheless, nine samples from the North-Lithuanian (Haanja) zone have a weighted mean age of 13.0 ± 0.8 ^{10}Be ka yr, close to radiocarbon and varvochronological measurements (Rinterknecht et al. 2006).

After the decision of the INQUA Congress in Paris in 1969 the base of Holocene Series is broadly accepted with an age of 10 000 radiocarbon years (some 11 500 calibrated years) what is very convenient in the global scale. Some time ago, ICS Subcomission on Quaternary Stratigraphy recommended to define the boundary at a depth of 1,492.3 m in the GRIP ice core from Greenland, reflecting the first signs of climatic warming at the end of Younger Dryas/Greenland Stadial 1 cold phase (Gibbard et al. 2008). The age of the boundary, based on multiparameter annual layer counting is 11.734 cal radiocarbon years before AD 2000, what is difficult to use in practice. The mentioned age is also rather disputable. It means that for the Baltic States the old boundary is much more preferable.

In conclusion, we can say that despite the great number of analyses and publications devoted to the Quaternary stratigraphy and marked improvements in study methods, many problems of topical interest have not been solved yet, especially due to the lack of good direct dating methods. Even the application of the most modern physical dating methods could not help improve the existing stratigraphical charts.

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DEVELOPMENT OF PALAEOLAKE HALJALA

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A 4.5 m thick sediment sequence from Haljala (59°25′27′′N, 26°17′42′′E; Fig. 1) at an elevation of 67.4 m a.s.l provided information on palaeoenvironmental changes in time range between 11 300 and 13 800 cal yr BP based on pollen record, AMS ^{14}C dates, sediment composition, plant macrofossils and ostracods. New proxies enabled to differentiate five main environmental stages, which have been tentatively related to the event stratigraphy of the Last Termination in the North Atlantic region (Lowe et al. 2008) and confirmed that ice cover on the Pandivere Upland started to perish already about 13 800 cal yr BP.

The basal part of studied sequence between 500 and 330 cm was represented by fine-grained and clayey silt and covered a time span 13 800–13 300 cal yr BP. This portion of sediments was deposited in a large proglacial lake during the generally warm middle part of the Allerød–Bølling GI-1c, supported by high organic matter (OM) content, peaking around 13 400 cal yr BP (Fig. 2). Decrease in arboreal pollen accumulation rate (PAR) around 13 600–13 700 and 13 100–13 300 cal yr BP referred to a short climatic deterioration within the Allerød interstadial. Between 13 300 and 13 100 cal yr BP fine-grained silt deposited with lower OM values than during the previous 500 years. Low primary pollen accumulation and OM content with high secondary pollen such as *Picea*, *Alnus*, *Corylus* and *Ulmus* percentages suggested climate cooling.

Between 13 100 and 12 850 cal yr BP fine-grained silt deposited with rather high OM matter reaching its late glacial maximum (7.3 %; Fig. 2). During this time span PAR of trees, shrubs and herb increased drastically confirming a short climate amelioration and establishment of pine-birch woods. This 250 year long warming could be synchronized with warmer climate during the later part of Allerød (GI-1a; Lowe et al. 2008).

Climate deterioration in the Younger Dryas (GS-1) was recorded from the reduction of tree pollen abundance and flourishing of cold-tolerant species, such as *Artemisia*, Chenopodiaceae, Cyperaceae which evidence a significant reduction of Sub-Arctic woodlands and replacement by grass–shrub tundra. About 13 000 cal yr BP the Haljala sedimentation basin isolated from the large proglacial lake forming first a narrow sound which finally isolated about 12900–12800 cal yr BP after the spit between Tatruse and Vanamõisa bedrock hills was shaped (Fig. 1).

Considerable change in the vegetation community occurred at about 11 700 cal yr BP with the start of Holocene warming. The increased representation of all forest taxa, first of all *Betula*, suggested that forest area expanded again, characteristic to the beginning of Holocene. Some delay occurred in vegetation response to ameliorated climate as it also was described on the Karelian Isthmus (Subetto et al. 2002; Wohlfarth et al. 2002) and motivated with the cold water of the Baltic Sea and possible increase of anticyclone circulation due to presence of Scandinavian Ice sheet remnants.

Acknowledgements: This research was supported by ETF Grants 6736 and 7029 and by the SYNTHESYS Project under the FP6 “Structure the European Research Area” Programme.

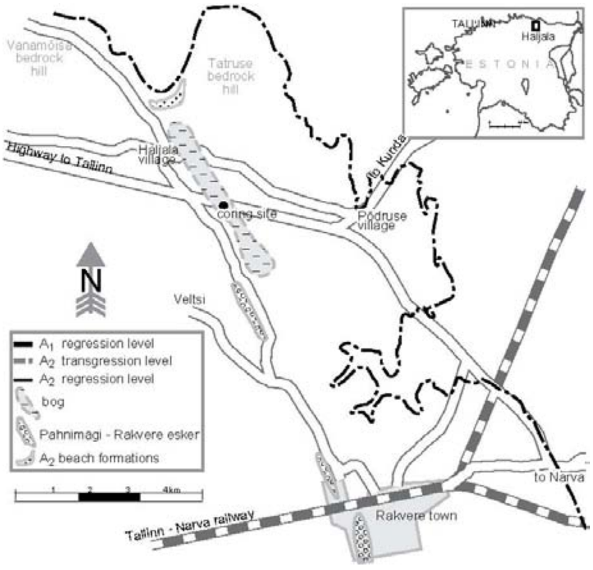

Fig. 1. Location of study area.

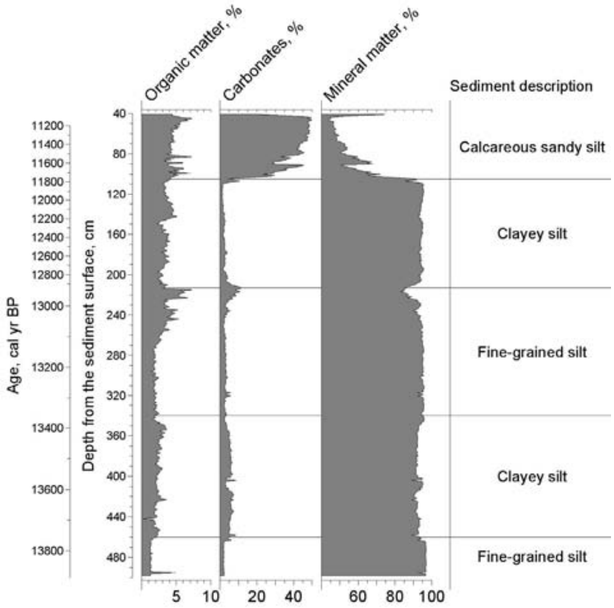

Fig. 2. Organic, carbonate and mineral matter content of studied sequence.

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A GLACIAL DYNAMIC STUDY OF APRIKI TONGUE: IMPLICATIONS FOR DEGLACIATION HISTORY

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In this study we evaluate glacial dynamics and subglacial processes of Apriķi glacial tongue in Western Latvia, during Linkuva deglaciation phase of the last Scandinavian ice sheet (Fig. 1). Geological setting of the study area consists of up to 60 m thick loose Pleistocene sedimentary sequence, which is thinning out in the direction of the glacial movement. This study comprises mapping and interpretation of three outcrop sections corresponding to the lateral margins and central zone of the glacial tongue (Fig. 1) as well as mapping and interpreting geomorphological glacial features from the digital terrain model and aero photographs. The geological structure and geomorphological features are interpreted to reconstruct the glacier bed conditions and ice movement patterns in this area.

The extent of Apriķi glacial tongue is marked by characteristical set of glacial landforms – marginal-shear moraines, drumlin-like landforms in the central part and marginal ridge at the distal end (Fig. 1).

Unidirectional thrusting and folding of subglacial strata corresponds to marginal shear moraines, situated at the margins of the Apriķi glacial tongue. These features were formed due to side drag stresses, generated between ice masses with differential velocity. On the topographical surface marginal shear moraines are expressed as linear ridges, stretching parallel to the glacial flow direction. Central part of the Apriķi glacial tongue dynamics is characterized by the decoupled sliding of the glacier, and vertical diapirism of the deeper layers, probably induced by the rapid changes of subglacial porewater pressure. Further down the glacial tongue Pleistocene sediments are thinning out to some 15–20 m in thickness, as well as bedrock geology changes from Devonian sandstones and siltstones to dolomites. This change coexists with onset of the drumlins on the Apriķi glacial tongue bed, suggesting that processes at the bed Apriķi glacial tongue was mostly governed by the hydrologic properties of glacial bed.

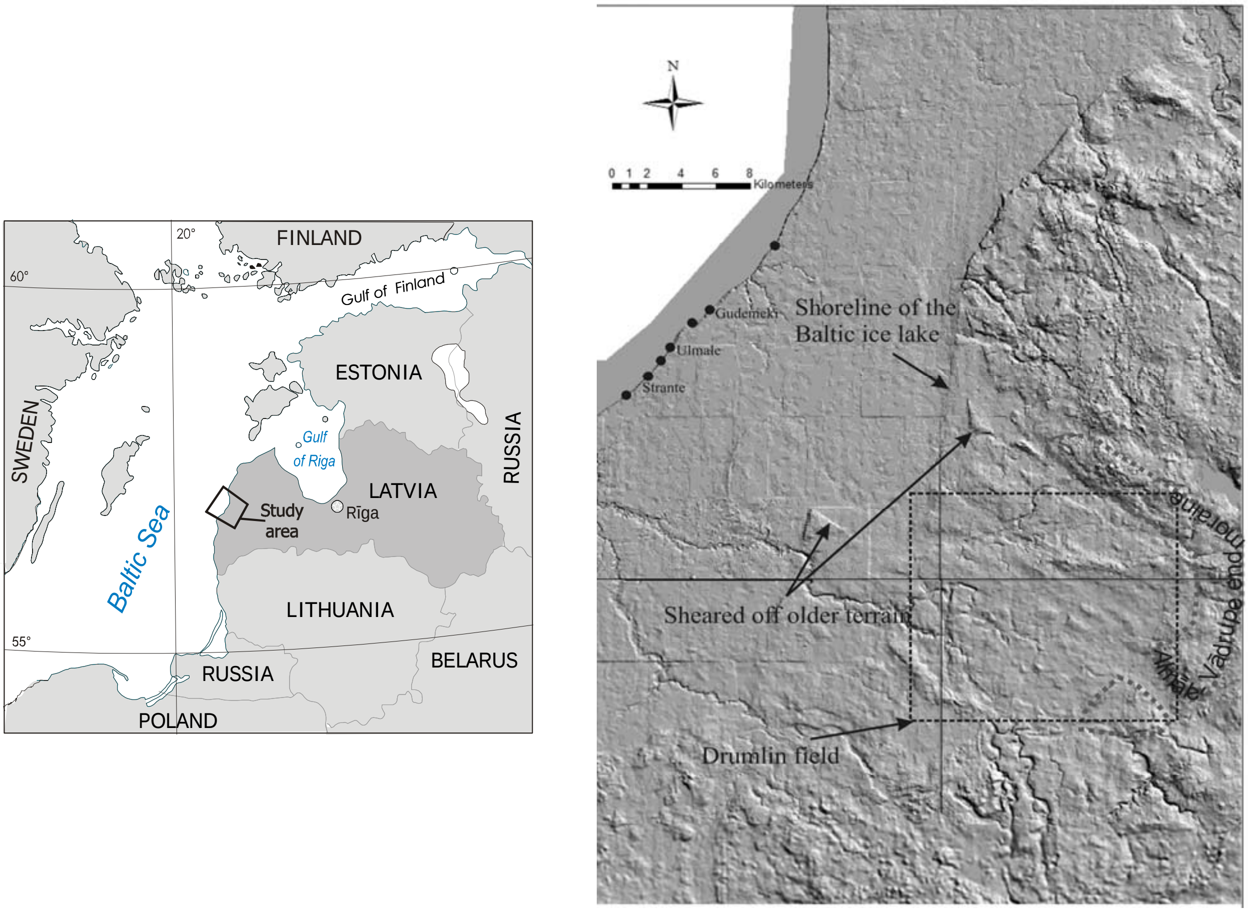

Fig. 1. A – Position of the study area. B – Digital terrain model of the Apriķi glacial tongue area. Vertical scale is exxagerated 3 times. Dots represent key sections at the Baltic sea coast. End moraine of the Apriķi glacial tongue is indicated with the dotted line.

We suggest that Apriķi glacial tongue formed in a surge-type event: development of rapid ice flow in largely stagnant ice as a result of either renewed ice accumulation or loss of basal sediment strength due to rise of relative pore water pressure as the ice gradually melted away. The ice flow drained ice from surrounding area inhibiting the re-activation of whole ice body. It is suggested that several such local rapid ice flows formed during delectation of last Scandinavian glaciations in Western Latvia.

RE-EXAMINATION AND USABILITY OF DIFFERENT TILL GEOCHEMICAL DATASETS – EXAMPLES OF MINERAL POTENTIAL MAPPING IN NORTHERN FINLAND

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Till geochemistry combined with surficial geology has been used as a practical mineral exploration tool in glaciated areas since 1950's. The golden age of large, nation-wide sampling projects was in the 1970's and 1980's. For example, a sampling project with the sampling density of ¼ km^2 (regional till geochemistry: Salminen 1995), carried out during the 1980's forms good foundation for the geochemical examination of background values and mineral potentiality in Finland. Since those days, increased knowledge and better understanding of glacial deposits gained with new methodologies has given possibilities to direct mineral potential mapping and exploration, and also re-examination of previous results and datasets in a new way. Furthermore, a multidisplinary approach using various exploration methods, better chemical analysing methods, and GIS-based processing allow better interpretation and analysis of the results.

Re-examination and usability of different till geochemical datasets have been started to analyze in 2008 at the Geological Survey of Finland. The work is done in the project 'Research of surficial deposits, their geochemistry and stratigraphy, and it will last several years. The main purpose is to gather all the geochemical datasets from the certain areas together and analyze them with GIS and statistical methods. With the multidisplinary approach, it is possible to re-analyze and compare geochemical datasets of different ages, densities and variable methodologies to form new interpretations of their usability in the framework of modern geological knowledge. The work has been started with the datasets of samples taken by percussion drill. After them, the work will be continued to reinterpretation of analysis results of samples from test pits, stream sediments, water, and heavy minerals.

To study the new till geochemical interpretations, we introduce two study areas in northern Finland. The first study area is the Enontekiö region in the fell area of northwestern Finnish Lapland, and the second is the Meltaus region in central Lapland. In those areas glacial overburden is composed of till which thickness varies from three to five meters but in the low land areas even from ten to twenty meters. Both areas were included in the sampling areas of the northern Fennoscandia geochemical mapping project (i.e. Nordkalott project area; Bölviken et al. 1986) and the regional geochemical mapping project. The Enontekiö region has also been examined in active exploration projects during the last decades. Its bedrock is composed of mafic volcanic and sedimentary rocks with schists, which are similar to the bedrock of the mine district in the Central Lapland Greenstone Belt. Prospectivity related to nickel and other base metals together with gold exploration is expected. The Meltaus region, instead, is located in the Central Lapland Granitoid Belt where, in spite of low ore potentiality, high gold anomalies are found.

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TILL GEOCHEMISTRY AS AN INDICATOR OF GLACIAL TRANSPORT DISTANCE AND DEPOSITION OF MORAINE FORMATIONS IN THE CENTRAL PART OF SCANDINAVIAN ICE SHEET

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Surficial geology, till geochemistry and heavy mineral investigation are used as practical exploration tools in the glaciated areas nearly one hundred of years. Since 1950's geochemical methods have been in a main role in mineral potential mapping and exploration. Till, as a sampling media is very useful due to glacigenic nature and the composition of fresh bedrock surface, weathered bedrock and older sediments. Till debris and fragments are always some distance derived from the source(s) and have dispersed to the direction of ice-flow, giving also a larger and more homogenized indication of source than the bedrock itself.

Secondary dispersion is influenced by many different factors. For example, geology, topography and subglacial conditions with ice-mass variations have great effect on the till deposition and the moraine formation. Deposition processes and transport distance of till vary in different parts of depositional environments relating to glacier body. Subglacial formations of advancing ice differ from the passive, melting-ice moraines. Furthermore, the active-ice formations like ribbed moraines, drumlins and flutings vary also considerably from each other. Till deposits in the ice-divide zone and in the interlobate areas have also own characteristics.

The determination of transport distance of till is conventionally done by using fabric and striae analyses, stone countings, grain size analyses, surficial boulder countings, i.e. till stratigraphical analyses. Latest surficial geological and mineral potentiality studies in northern Finland (Sarala 2005, Sarala et al. 2007, Sarala & Ojala 2008) have proven that till geochemistry is an effective way to estimate transport distances and deposition processes of mineralized material in till. Heavy mineral investigations are supporting till geochemical studies.

Several case studies in southern Finnish Lapland show the extremely strong glacial quarrying activity and short transportation of till from the underlying bedrock to the top of ribbed moraine ridges. It is seen in till geochemistry as sharp and strictly bordered metal anomalies in the areas where mineralizations exist in the bedrock. The distance of the highest anomalies can be only some tens of meters from the source like in the Petäjävaara and Kuohunki targets. Similar short transportation is seen in some targets in the central Lapland area. One of the known gold mineralizations in the Petäjäselkä, Kittilä cause the highest anomalies in the bottom of two to three meters thick till cover only after a few meters transportation. Indication of the same enrichment of gold is seen in the topmost part of till cover about 15 to 20 m down-ice from the source. Also somewhat longer (some tens of meters to some hundreds of meters) dispersion patterns in till geochemistry are common in many target areas.

In the drumlin and fluting areas, the deposition history of the formations are more complicated and transport distances much longer (500 m to some kilometres) than earlier deposits. Streamlined moraine formations are common in the areas of actively streamed glaciers but due to warm-based, watery subglacial conditions, the transport distances of till have been long. In the till geochemistry this is seen as a low content of elements in till, which is a consequence of large source areas under warm-based glaciers. The same situation is in the areas of ice-divide zone and interlobate areas, where the effect of passive ice has caused low glacial erosion, and in places, only a redeposition of pre-existing glacigenic deposits. Till in the uppermost layers can also be formed of long-distance debris and rounded rock fragments that have released from the melting but at the last phase, non-erosive glacier. In those cases the contents of elements have diluted to the background level.

The same phenomenon is seen in heavy mineral investigations. In the areas of short and effective glacial erosion and followed transportation, fresh heavy mineral grains, like sulphide minerals and gold nuggets, have been preserved. Instead, in the areas of till with mature composition (rounded, small-size rock fragments and high amount of the fine fraction of till) the preservation of fresh heavy minerals is weak, and the indication of underlying bedrock is poor.

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DEVELOPMENT OF DUBICIAI GLACIOFLUVIAL BASIN AND IT'S INFLUENCE FOR SOUTH LITHUANIA DRAINAGE SYSTEM

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The Dubičiai glaciofluvial basin is in Lithania – Belorussian boundary (South of Lithuania) (Fig. 1).

The Dubičiai glaciofluvial basin was research by geomorphological, lithological and cartographic methods. Such methods let estimated old water surface level. Leveling of wide bench and upper terraces show, that rising of water level was gradual. It's was determined by large basin area and small-scale water inflow, which was determined by slowly climate changes. The high water level created two benches: lower 2 meters deep and lower about one meter deeper. Silt and other lacustrine sediments was covered deeper bench.

The stable basin level created good condition for sedimentation of saprophel, freshwater limestone, shells, and plant residuals. In shallow bench accumulate varied terrigenic sediments. Where the water level was arising, the sedimentation zones approached to shoreline. The sand and gravel covered silt sediments. Low amplitude water oscillation didn't influence for deep-water sediments. Where water level fall 5–10 m, accumulation condition strongly exchange.

Formation of glaciolacustrine basin and its sedimentation processes was determined by terrain deglaciation. In Frankfurt Stage was formed glaciofluvial thermokarst terrace (before 20 000 year). Lower terrace (140 m NN) was formed by glaciolacustrine basin in Dryas – Alerod epochs (before 12 000 – 10 000 year). Lower level was formed in Boreal – Alerod time (before 10 000 – 6 000 year). In Subatlantic epoch (before 5 000 year) was formed lowest terrace level – Ūla terrace. 150–100 years ago the Dubičiai basin almost fully outflows. Latest drainage stage was in 1958–1959 years, when wide drainage project was executed (Fig. 2) (Baltrūnas 2002, Kabailienė 2006, Stančikaitė et al. 2002, Švedas et al. 2004).

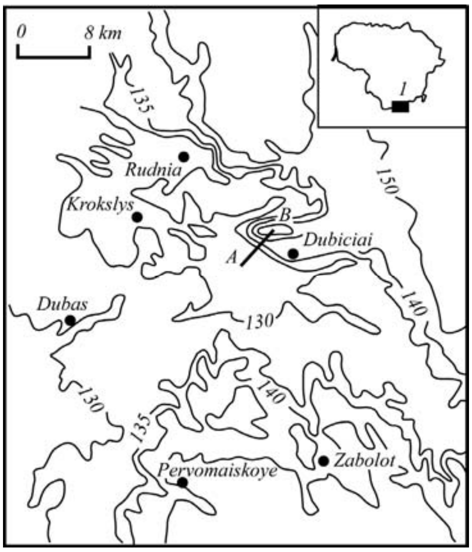

Fig. 1. Eastern part of Dubiciai basin. 1 – investigation area

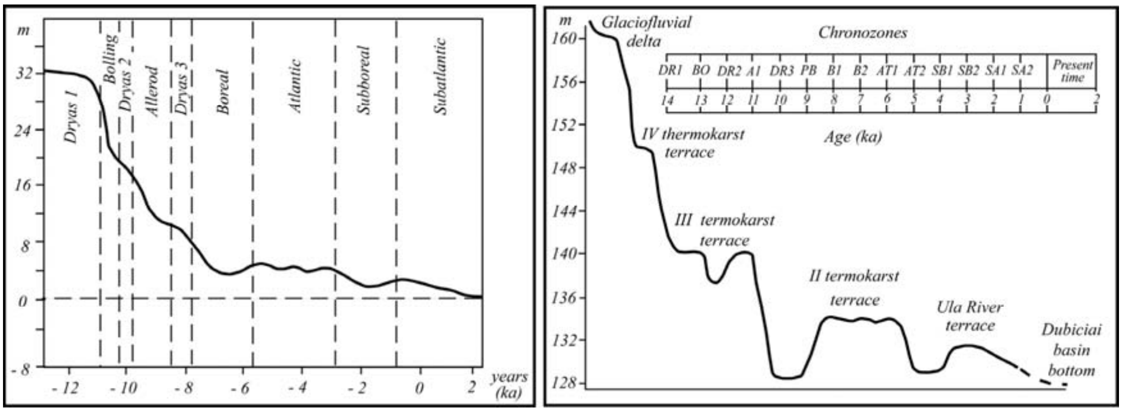

Fig. 2. Water level changes in Dubiciai Basin during Late Pleistocene and Holocene.

Intense relief transformation was in end of late Pleistocene and in early Holocene. Largest relief changed by thermokarst processes, which was beginning in Alerod and continued by all Boreal. Much deep and large thermokarst relief form as lakes rines was formed. Some thermokars lakes as Dumblys, Dumblelis, Grikis, Katišius and Tabelis are in Dubičiai depression along end moraine ridges.

Beginning Preboreal climate all time becoming warm and moist. Especially warm and wet it was during late Boreal, Atlantic and Subboreal. The climate changes become the water level changes in lakes and basins. The low water level was during all Preboreal and early Boreal. In this time water level was lowest by all Holocene epoch. The warmer climate formed good conditions for peat, gytia and freshwater limestone accumulation. In another side, it's decreased terrigenic material accumulation. During later Boreal, Atlantic and early Subboreal was intensive accumulation of peat, gytia and freshwater limestone in Dubičiai basin. In Subatlantis epoch dominated peat accumulation and wetland formation.

The latest stage of Dubičiai basin was beginning in middle XIX century. During 50 year area of basin decrease 10 time: from 221 ha in 1850 year to 20 ha in 1900 year. In first half of XX century Ūla River deepened its longitudinal profile. Its inspired capping processes: upper part of Katra River became to Ūla catchment. Catchment transformation determined shallow lake

(Matarai, Pelesa, Duba) drainage. The existence Dubičiai basin was finished in 1958 – 1959 year when wide drainage project was executed. It's decided that decrease water level in Tabalis and Stojai lakes (Eitmanavičienė & Endzinas 1977).

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SIGNATURE OF THE VISTULA ICE STREAM IN NORTHERN POLAND DURING THE LATE WEICHSELIAN GLACIATION

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Glacial dynamics of the last Scandinavian Ice Sheet in northern Poland was influenced by a few second-rank ice streams, which were fed by the Baltic Ice Stream, functioning along the Baltic Sea basin (Punkari, 1993; Boulton et al., 2001). The Vistula ice stream was one of the major dynamic elements of the ice sheet periphery, controlling the spatial and temporal variability of regional-scale ice masses (Marks, 2002, 2005; Wysota, 2002). The soft-bedded and land based ice stream moved along the Vistula Valley depression towards the ice sheet margin, where it formed the broad Vistula ice lobe. The glacial sequence of the Vistula lobe reveals two ice advances of varied extent during the Late Weichselian (Wysota 2002, 2007; Wysota et al. 2002, 2009).

Convincing geomorphological and geological records supporting fast ice flow in the Vistula lobe were found. Geomorphological features include: the trough-shaped exaration depression, low relief till plains and streamlined landforms. Geological signatures of rapid ice movement are: unconformities under tills of the Vistula lobe, uniform thickness and composition of subglacial till sheets, deformation till facies, boulder pavements with striae and ploughing marks. It is accepted that fast flow of the Vistula ice stream was facilitated by combination of bed deformation and enhanced basal sliding.

The research has been financially supported by the Polish Ministry of Science and Higher Education on the basis of the project no. N N306 316835.

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ON THE CHRONOLOGY OF THE NATURAL EVENTS DURING THE POOZERJE (WEICHSELIAN) GLACIATION ON THE TERRAINS OF BELARUS

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Based on modern data from the palynological analyses of the Poozerje glacial complex (quantity of discharge phases of the development of vegetation in different time frames) and the data on the variations of the Late Pleistocene climate from the oxygen isotopic curves, the Poozerje glacial period within the limits of the Balorussian terrain, is at present compared with isotope stages 2–4 (10300–70000 years ago). Duration of the Murava (Eemian) interglaciation with the three climatic optimums lasted not less than 40–50 ka years and covers the whole isotope stage 5, and not just one part (5-e) of it. On the duration (about 60 ka years) the Weichselian glaciation practically does not succumb to other graciations in the Pleistocene. Duration of the glaciations was as follows (in thousand years): Narew – 30, Servech – 50, Berezino – 15–20, Eselevo – 20, Yachny – 50, Dniepr – 55, Sohz – 15.

Maximum areal distribution of continental ice during the Pleistocene occurred during the Dniepr glaciation (OIS 8) when glaciers completely covered all terrain of Belarus and reached middle of Ukraine. Maximum cooling happened during the Poozerje glaciation (OIS 2-4). The latter judgement is not recognized by all researchers because the ice reached only the northern part of Belarus during the Poozerje glaciation.

The main temperature decrease took place in the second half of the Poozerje glaciation — during the maximum of the Orscha megas-tage approximately 17 ka ago (beginning of OIS 2). Poozerje glaciation resulted the formation of glacial horizon from till and fluvioglacial deposits within the limits of glaciation, and alluvial, limnic and mire deposits on the terrains outside of the glaciated area (in thickness of up to 70 m). In the structure of that glacial horizon Kulakovo, Dvina and Naroch sub-horizons have been determined, reflecting the three main epochs (early, middle and late) of the Poozerje glaciation (Table). The divisions have a different duration depending on the phases of the development of vegetation, described in the palynological diagrams.

Poozerje glaciation clearly enough was exhibited in the Northern hemisphere in the terrain of Eurasia and North America, it corresponds to the sequences at Lashamp (43–45 ka years ago), Mono (24–25 ka years ago), Goteborg (12–14 ka years ago), Würm, Würm-1-2, Würm-2, Wisla-1-2-3, North-Polish, II-Warsaw, Tubant horizon in the Netherlands; Kalinin, Valdai, Olonec, Leningrad, Ostashkov horizons in Russia, Nyamunas in Lithuania; Udai, Vitachev, Bug, Dofinov and Prichernomorje in Ukraine.

Our ancestors could feel on themselves the influence of this natural phenomenon, under which they gradually moved from the south on the north in the process of the redemption of vast terrains from the cover continental ice. Development of the environment at all stages of the Poozerje glaciation and the Holocene is known for us in greater detail, than that of during other epochs of the Pleistocene. Available numerous palaeontological and the geochronological data allows the researchers to present the climatic megacycle – Poozerje glaciation – Holocene interglacial, as the measurement of the standard for the forecast of change of a climate hereafter.

Table. Chronology and stratigraphy of the Poozerje glaciation (PZ) in Belarus (by Ya. K. Yelovicheva)

Isot. sta-ges	Sub-divi-sions	Periods		Phases of evolution of the vegetation		*Geochro-nological interval
1	Holocene interglacial			HL		Modern
PZ 2	Late — Naroch	Dryas-III stage		DR-III	pz-f-6	10300 л.н.
		Allerod interstage		AL	pz-f-5	10800
		Dryas-II stage		DR-II	pz-f-4	11800
		Belling interstage		BL	pz-f-3	12300
		Dryas-I stage		DR-I	pz-f-2	12700
		Raunis interstage		RN	pz-f-1	13000
	Middle — Dvina	Orsha megastage (maximum)		OR	pz-s-14	13900
		Usvyach preglacial suite		UV	pz-s-13	17000
PZ 3		Megainterstage	Borisov interstage	BR	pz-s-12	26000
			Michalinovo stage	MH	pz-s-11	28000
			Schapurovo interstage	CH	pz-s-10a-b	30000
			Rogachev stage	RG	pz-s-9a-e	35000
			Turov interstage	TR	pz-s-8	39000
PZ 4		Megastage	Mezin megastage	MZ	pz-s-7	44000
	Early Kulakovo	Megainterstage	Polotsk interstage	PL	pz-s-6a-c	49000
			Sloboda stage	SL	pz-s-5a-d	52000
			Suraz interstage	SR	pz-s-4a-c	55000
			Mirogotschi stage	MRG	pz-s-3	58000
			Cherikov interstage	CR	pz-s-2a-b	60000
		Megastage	West-Dvina megastage-2	ZD-2	pz-s-1-c	6400
			Black-bereg interstage	CHB	pz-s-1-b	66000
			West-Dvina megastage-1	ZD-1	pz-s-1-a	68000
5	Murava interglacial			MR	mr-1-10	70000–110000

s – start and f-final phases of the vegetation
* dates of the high bounds of subdivisions

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OUTLINES OF THE QUATERNARY GEOLOGY OF THE REGION

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Estonia (45215 km^2) and Latvia (64589 km^2) are located on the northwestern part of the East-European Platform and the region is bordered on the north by the southern slope of the Baltic Shield. In the subsurface, the Precambrian basement of crystalline rocks and the succeeding Upper Proterozoic and Palaeozoic sedimentary rocks underlie the territory of Estonia and Latvia. The total thickness of the sedimentary bedrock increases from about 100 m in the northern Estonia to nearly 2000 m in the south-western Latvia (Raukas & Teedumäe 1987, Dreimanis & Kārklinš 1997).

Quaternary deposits of various thickness and age cover almost all Estonian territory (with the exception of restricted bedrock outcrop areas) and all of Latvia. In northern and north-western Estonia, the thickness of Quaternary deposits usually does not exceed 5 m. In Estonia the Quaternary deposits are at their thickest (50–200 m) in the Fore-Clint Lowland at the coast of the Gulf of Finland, in Saadjärve Drumlin Field, in the uplands of South-East Estonia (Otepää and Haanja), and in the buried valleys (207 m as a maximum at Keskküla in south-western Estonia) (Kajak 1995).

The average thickness of the Quaternary deposits in Latvia is 5–20 m in the lowlands and 40–60 m in the highlands. Locally, along the southern, south-western and south-eastern sides of the bedrock elevations, e.g. in the Vidzeme Upland, and in east-central Latvia, between Cesis and Madona the thickness of the Quaternary deposits may be up to 200 m. The greatest thickness of the Quaternary deposits (310 m) occurs in the Aknīste buried valley in south-eastern Latvia (Zelčs & Markots 2004).

Most of the Quaternary deposits in Estonia (95%) and Latvia are glacigenic and of Pleistocene age. The landscape of the region bears distinct traces of glacial activity, and commonly the cores of positive glacial landforms consist of glaciotectonically deformed sediments. The highest points in Estonia and Latvia are respectively the Suur Munamägi (318 m a.s.l.) in the Haanja Upland, and the Gaiziņkalns (312 m a.s.l.) in the Vidzeme Upland.

In Estonia five till beds, often of great thickness, are more or less continually traceable. Only in few places the till beds are separated from each other by Holsteinian (Karuküla) and Eemian (Prangli) interglacial or interstadial deposits (Raukas et al. 2004). In Latvia, deposits of four glaciations and three interglacials have been identified (Zelčs & Markots 2004). In both countries the age of the till beds, if possible, is determined by their stratigraphical position with respect to the interglacial deposits, or in many cases by their bedding position with respect to the till beds with known age. Strong erosion by subsequent glaciations has destroyed much of the evidence of the Early and Middle Pleistocene glaciations in the region. Based on present knowledge the Early Pleistocene deposits are totally absent from Estonia and the Middle Pleistocene sequence is incomplete here. Deposits of Elsterian (Sangaste) glaciation are the oldest in Estonia. In Latvia, the oldest till (Latgale Formation) probably represents an Early Pleistocene glaciation (Dreimanis & Kārklinš 1997) (see Table).

Time transgressive and regressive processes of glacial accumulation, with particular importance of selective glacial erosion, and proglacial melt water activity, have resulted in the formation of lowlands, while proglacial and subglacial accumulation and glaciotectonics predominated in the glacial uplands. In this sector (south-eastern) the maximum extent of the Weichselian glaciation (LGM) occurred at ca 18.3 cal. ^{14}C ka BP in Lithuania (Rinterknecht et al. 2008) and at ca 18 cal. ^{14}C ka BP southeast of the Onega basin (Lunkka et al. 2001) in

Russian Karelia. Morphological evidence persisted from large ice streams in length of 100 to 300 km that drained the ice sheet after the LGM (Fig. 1). These ice streams destroyed the record of possible older ice streams at least in lowland areas. The area between LGM and the Baltic Sea did deglaciate very rapidly, approximately within 5–6 ka, as the last morphologically well-expressed marginal formation before the Salpausselkä I – the Palivere ice marginal zone – did form ca 12.7 cal. ^{14}C ka BP (Kalm 2006, Raukas 2009).

NE Europe & Scandinavia			OIS	Estonia		Latvia	
Holocene			1	Holocene		Holocene	
Weichselian	Late Weichselian		2	Järva	Võrtsjärve	Latvian	Zemgale
Weichselian	Middle Weichselian	Denekamp Hengelo Moershoofd Oerel	3-4	Järva	Savala	Latvian	Vidzeme Kursa
Weichselian	Early Weichselian	Odderade Brørup	5a-d	Järva	Valgjärve Kelnase	Latvian	Early Latvian
Eemian			5e	Prangli		Felicianovian	
Saalian	Late Saalian (Warthe & Drenthe)		6	Ugandi	Late Ugandi Middle Ugandi Early Ugandi	Kurzeme	
Saalian	Early Saalian		6-8	Ugandi		Kurzeme	
Holsteinian			9 (11?)	Karuküla		Pulvernieki	
Elsterian				Sangaste		Letiža	
Cromerian complex						Židini	
Bavelian						Latgale	

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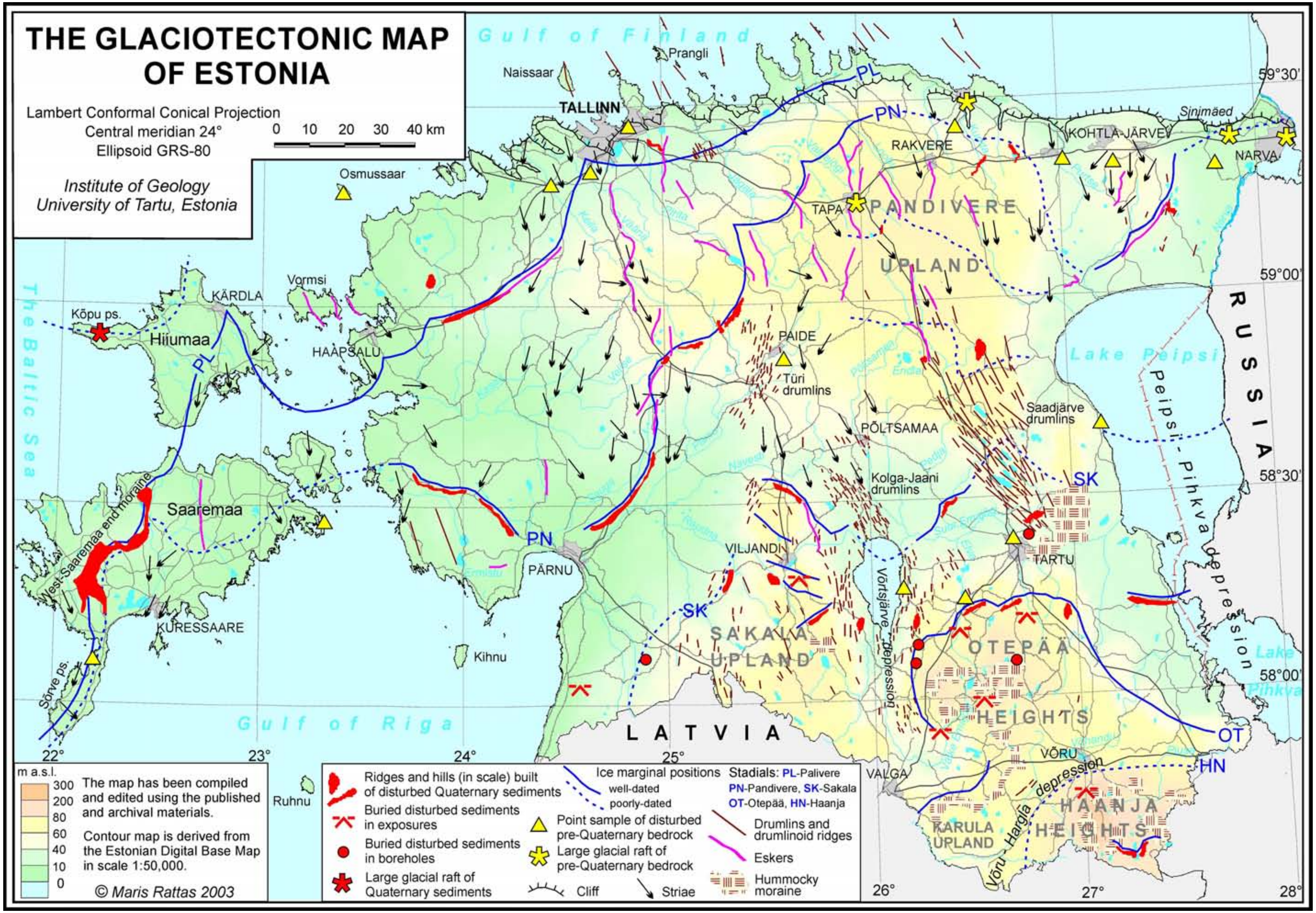
Fig. 1. DEM of southern Estonia and northern Latvia. White solid line marks the national border between Estonia and Latvia. Red lines indicate general orientation of subglacial bedforms; black lines denote major ice marginal positions.


THE GLACIOTECTONIC MAP OF ESTONIA
Lambert Conformal Conical Projection
Central meridian 24°
Ellipsoid GRS-80
0 10 20 30 40 km
Institute of Geology
University of Tartu, Estonia
Gulf of Finland
The Baltic Sea
Gulf of Riga
RUSSIA
LATVIA
Lake Peipsi
Lake Pihkva
Peipsi - Pihkva depression
Naissaar
Prangli
Osmussaar
TALLINN
Vormsi
KÄRDLA
Kõpu ps.
Hiiumaa
HAAPSALU
Saaremaa
West-Saaremaa end moraine
KURESSAARE
Sõrve ps.
Kihnu
Ruhnu
PÄRNU
PAIDE
Türi drumlins
PÕLTSAMAA
Kolga-Jaani drumlins
Saadjärve drumlins
TAPA
RAKVERE
PANDIVERE UPLAND
KOHTLA-JÄRVE
Sinimäed
NARVA
VILJANDI
SAKALA UPLAND
Võrtsjärve depression
TARTU
OTEPÄÄ HEIGHTS
VALGA
VÕRU
KARULA UPLAND
Võru - Hargla depression
HAANJA HEIGHTS
PL
PN
SK
OT
HN
59°30'
59°00'
58°30'
58°00'
22°
23°
24°
25°
26°
27°
28°
m a.s.l.
300
200
80
60
40
10
0
The map has been compiled and edited using the published and archival materials.
Contour map is derived from the Estonian Digital Base Map in scale 1:50,000.
© Maris Rattas 2003
Ridges and hills (in scale) built of disturbed Quaternary sediments
Buried disturbed sediments in exposures
Buried disturbed sediments in boreholes
Large glacial raft of Quaternary sediments
Ice marginal positions
well-dated
poorly-dated
Point sample of disturbed pre-Quaternary bedrock
Large glacial raft of pre-Quaternary bedrock
Cliff
Striae
Stadials: PL-Palivere
PN-Pandivere, SK-Sakala
OT-Otepää, HN-Haanja
Drumlins and drumlinoid ridges
Eskers
Hummocky moraine

STOP 1:

GLACIOTECTONIC DEFORMATIONS IN PREEDIKU GRAVEL PIT AT THE OTEPÄÄ ICE MARGINAL ZONE (LATE WEICHSELIAN)

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The Preediku gravel pit (58°13′26′′ N, 27°30′53′′ E) is located on the northern slope of the Otepää Upland, about 1 km up-glacier from the terminal margin of the Late Weichselian Otepää Stadial (Fig. 1A). The proximal slope of the Otepää Upland is characterized by slightly hummocky and undulating topography with more or less isometric low hillocks and depressions (some of them as kettle-holes). Thickness of the Pleistocene deposits varies from about 10 m at Nõo to about 60–70 m in the marginal ridges. Two parallel NW-SE stretching end moraine chains at Tamsa and Kambja (half-transparent blue lines in Fig 1A) mark here the Otepää ice marginal zone. Both ridge systems are ca 4 km long with the highest points up to 130 m a.s.l., and with relative height of about 30 m. The ridges are mainly composed of outwash gravel and sand, which here and there are graded up to lacustrine fine sand, silt and clayey rhythmites. All these aquatic sediments are overlain by till.

The Preediku gravel pit is located on the top of an isometric hillock at about 95 m a.s.l. (Fig. 1B). The hillock measures about 200 m in diameter and about 10 m in relative height. Thickness of the Pleistocene deposits is ca 30 m from which the meltwater deposits form up to 16 m thick bed. In general, the lithologic succession of the site comprises following facies: (1) the uppermost reddish-brown loamy till with pebbles; (2) lacustrine silt and clay rhythmites; (3) outwash sand and gravel; (4) basal lodgement till. The lower till layer has been described only in boreholes and is not well investigated.

The outwash facies occurs as an isometric lens-like bed laterally limited to the boundaries of the hillock. Outwash deposits consist of alternating series of cross-bedded sand and gravel layers, that are generally dipping to the southeast and south. Grain size of the outwash deposits varies from fine sand to coarse gravel and pebbles. The coarser material dominates in the proximal part of the hill, indicating that the flow to the fan was from the direction of ice-margin. The outwash complex is overlain by lacustrine sediments with rhythmic structure that were deposited in proglacial body of water. Thickness of individual laminae in the lacustrine sediments varies from millimetre to few centimetres. This succession of laminated lacustrine deposits may be subdivided into two complexes according to their colour – the lower grey and the upper brown complex. The colour of the sediments is probably related with the oxic conditions in the proglacial lake.

Up to 6 m high southern wall of the gravel pit exhibits a succession of glaciotectonically disturbed meltwater sediments, that are discordantly overlain by thin till layer (Fig. 1C). The meltwater deposits were deformed in course of smooth folding and dragging but without considerable shearing. Some folds are highly attenuated and have almost horizontal bedding, others have steeply dipping to vertical flanks. Orientation of folds shows that the direction of main glacial stress was from NW and WNW.

Obviously the main phase of glaciotectonic deformations took place after the deposition of meltwater- and glaciolacustrine sediments. The ice-contact fan at Preediku was formed along a short-term recessional position of the ice margin shortly after the glacier receded from the slopes of the Otepää Upland. A short-lived proglacial lake existed between the ice and the main ice marginal ridge, where the lacustrine sediments deposited. The deformation processes took place during a short and spatially limited re-advance of the ice margin. Glaciotectonic structures and the discordant till layer on top of the succession suggests subglacial surficial folding and dragging deformations beyond the ice margin.

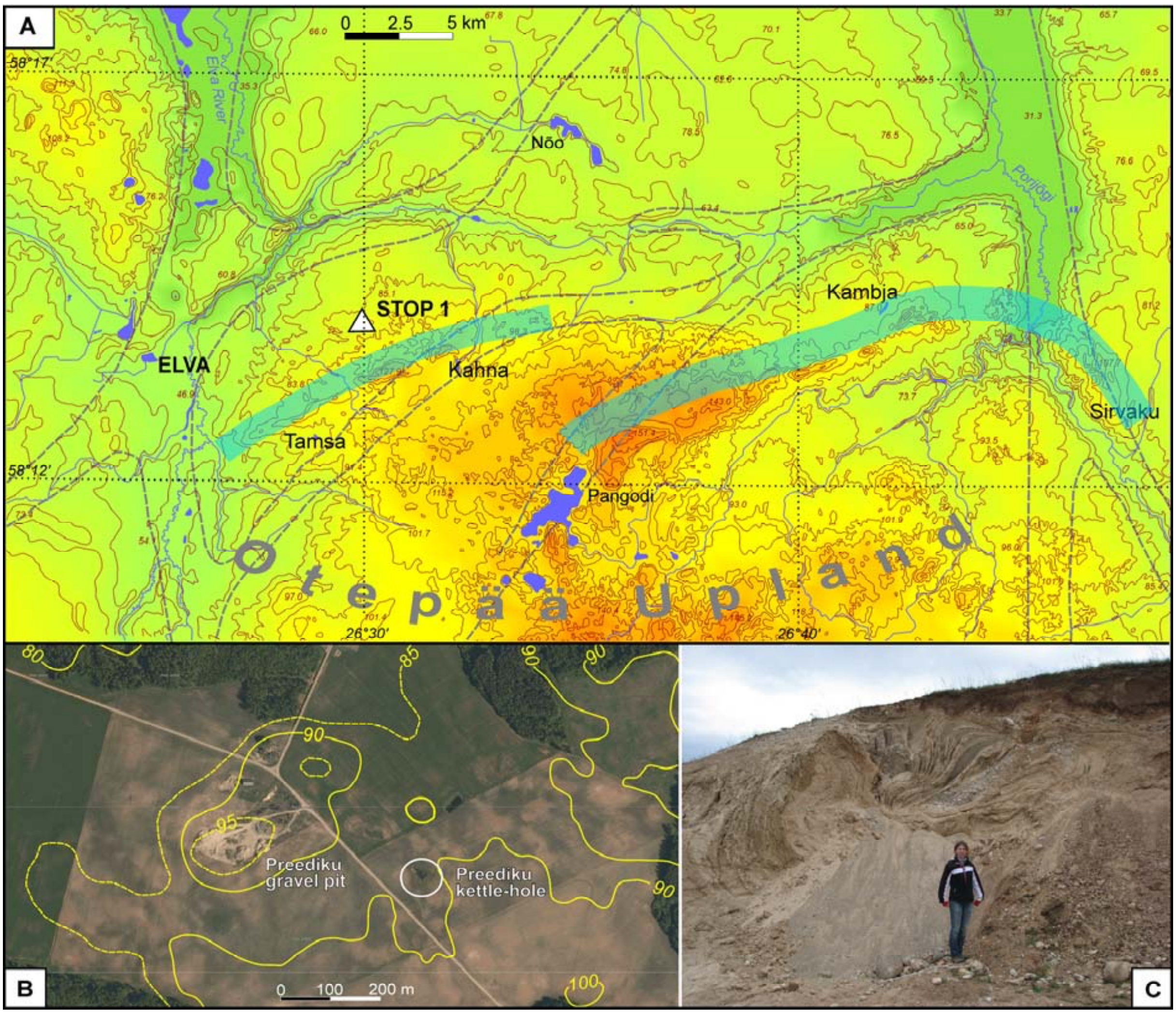

Fig. 1. A – Location of the Preediku site at the Otepää ice marginal zone (Late Weichselian, ca 14,5 ka BP). B – Slightly hummocky relief at Preediku, locations of the gravel pit and the kettle-hole are shown. C – Insight into the section of the deformed meltwater sediments in the Preediku gravel pit.

PREEDIKU KETTLE-HOLE

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Within 300 m from the Preediku gravel pit there is a kettle-hole (location 58º13'15,28''N, 26º30'7,8''E, see Fig. 1B), where sequence of Late Glacial (clayey silt) and Holocene (gyttja with tufa and peat; Fig. 2) sediments is opened in recent excavation. Till at the bottom of the kettle-hole is covered with thin layer (up to 23 cm) of fine gravel which is overlaid with presumably Late Glacial clayey silt and silty clay (up to 27 cm). Locally, where the surface of silty clay lies at lower elevations, it is covered by tufa-rich gyttja (< 30 cm). The thickest natural sediment layer (up to 1 m) is peat that covers silty clay and gyttja. Peat layer in turn is mantled with fill layer (disposed soil), presumably from the times of intensive amelioration in 20th century.

Recently three ^{14}C datings of sediments from kettle-hole have been accomplished: two from the bottom layer of the tufa-rich gyttja (organic matter and carbonate mollusc shells) and one from the bottom layer of peat. According to the datings, already at ca 11440 cal ^{14}C yrs BP began gyttja deposition in the small lake and the area had to be offloaded from a glacier much earlier. Shell fragments of the same tufa-gyttja layer yielded ca 800 yrs younger age (ca 10653 cal ^{14}C yrs BP), presumably indicating the onset of fertile living conditions for bivalves in the lake. In the tufa-rich gyttja, also an assemblage of ostracods was determined with species *Cyclocypris ovum, Limnocythere inopinata, Candona candida* and *Limnocytherina sanctipatricii*. Listed ostracod species indicate freshwater conditions in a cold-water shallow lake. Lacustrine phase of development in this kettle-hole ceased at ca 9600 cal. ^{14}C yrs BP when the first peat layer accumulated on top of the lake sediments.

From the preliminary data available we may conclude, that normal lake sedimentation in Preediku kettle-hole started (at 11.4 Ka BP), that is ca 3000 yrs after the glacier receded from the Otepää ice-marginal zone (at 14.5 Ka BP) and lasted until ca 9.6 Ka BP, when the lake turned into swamp.

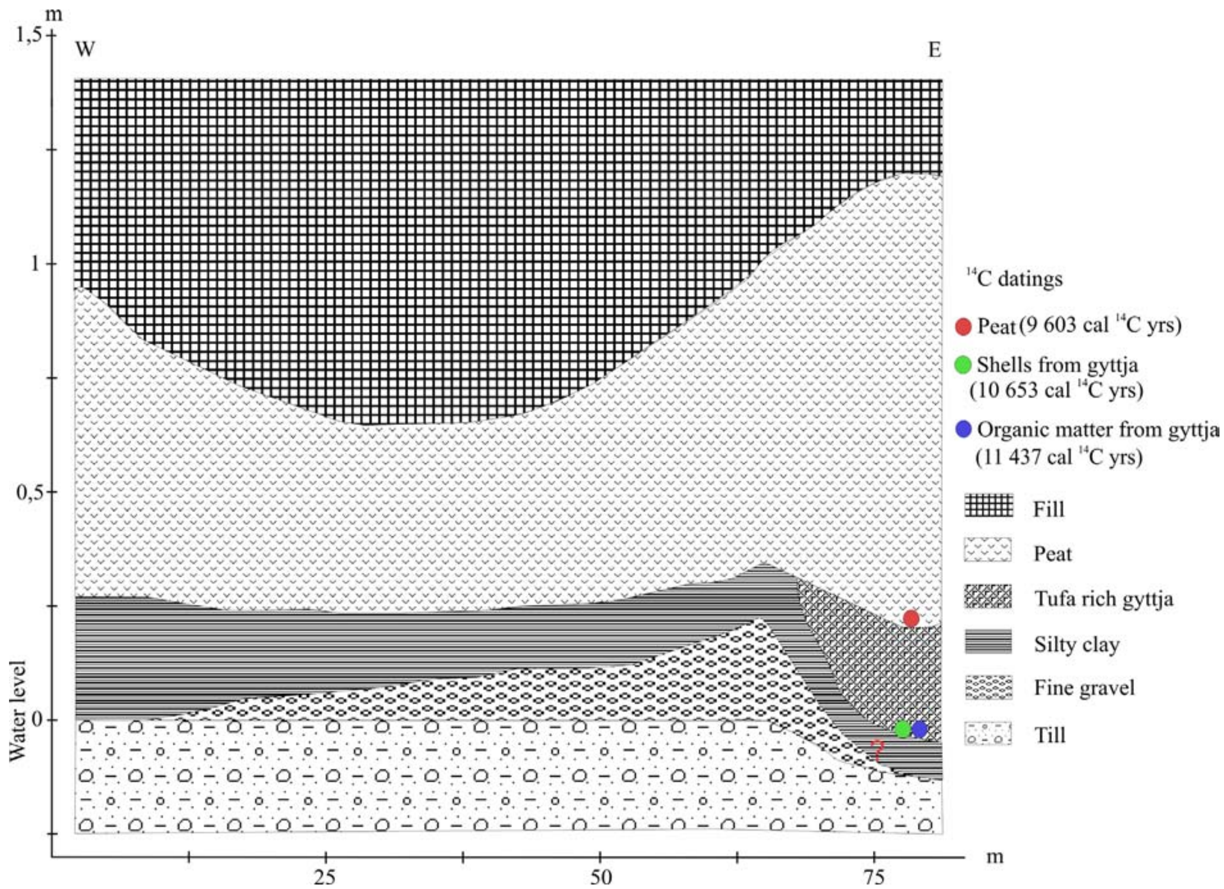

Fig. 2. Fragment from the cross section of the Preediku kettle-hole.

STOP 2:

GLACIOTECTONICALLY DISLOCATED EEMIAN INTERGLACIAL DEPOSITS AT RÕNGU

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Local farmer (Vaeva farmstead) in course of building a dug well discovered the Rõngu site of interglacial continental deposits (peat, gyttja, silty clay) in 1938. Shortly K. Orviku (1939) studied lithostratigraphy and P.Thomson (1939, 1941) biostratigraphy of the section and as a result, the Rõngu interglacial deposits were correlated with the Eemian deposits in Denmark and NW Germany as Jessen & Milthers (1928) described the latter.

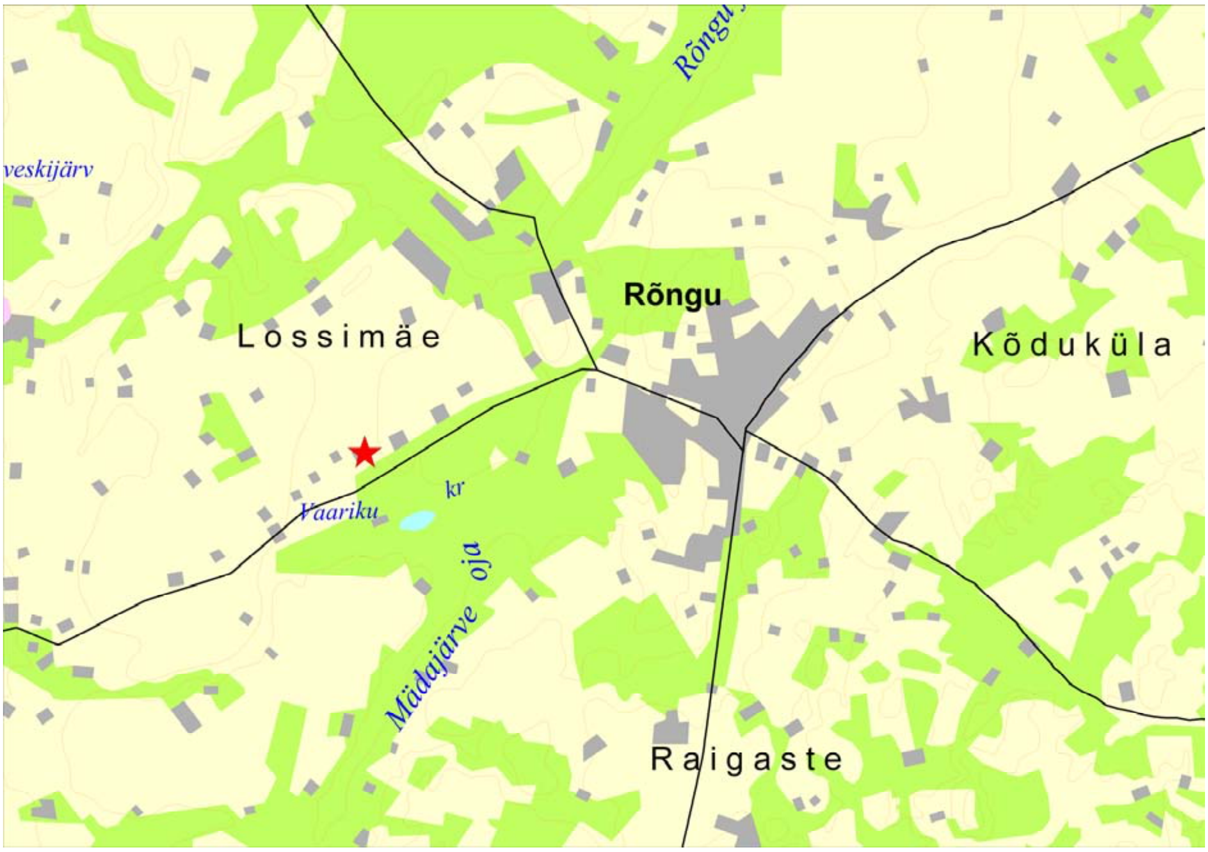

Fig. 1. Location of the Rõngu site of Eemian interglacial deposits.

Deformed bedding of the organic layers was noted already by K.Orviku (1939) and the deposits were thought to have been transported glacially from the Lake Võrtsjärv depression in northwest of the site. The reddish-brown till above and the grey till under the interglacial sediments were considered to be of Weichselian (Würm) and Middle-Pleistocene ages respectively. Peat and gyttja from Rõngu section were amongst the first interglacial deposits ^{14}C-dated at the Tartu Radiocarbon Dating Laboratory. The obtained ages were infinite (TA-45 and TA-46, both ≥30 000 ^{14}C yrs; in Liiva et al. 1966). In 1960-s and 1970-s the site was repeatedly investigated and number of new boreholes were made. As a consequence the distribution area of the

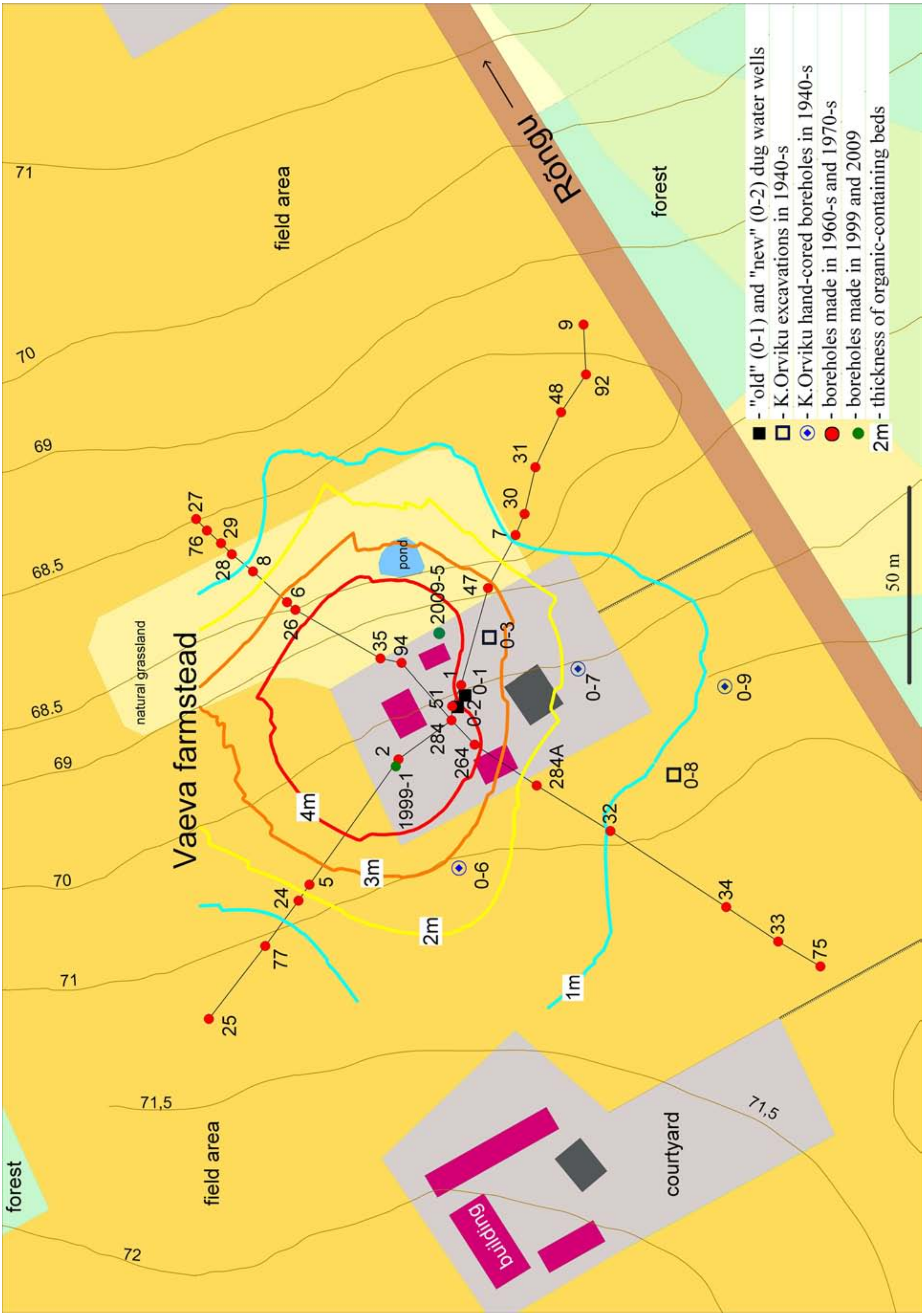

Fig. 2. Location of excavations and boreholes, and thickness of Eemian strata at Rõngu site.

interglacial deposits was determined (Fig. 2) and 10 main sedimentary units were identified in cross-section at the site (listed from the ground, with thickness of layers in brackets; see also Fig. 3):

1. Reddish brown clayey till (Late Weichselian till) (0–2 m);
2. Fluvial to lacustrine fine- to medium-grained sand, locally iron-stained (0–3.5 m);
3. Fen peat with pieces and lumps of compacted gyttja (0–1.5 m);
4. Gyttja, compacted, brecciated, with sand admixture, contains shell detritus and fragments of wood (0–3 m);
5. Medium to coarse sand with pieces of gyttja (0–1 m);
6. Grey pebbly till with fragments of compacted gyttja in its upper 0.5 m, considered to be of Late Saalian age (0–6 m);
7. Violet-grey till, considered to be of Early Saalian age (3–5 m);
8. Lacustrine clayey silt and sand, considered to be of Early Saalian age (7–15 m);
9. Grey till, considered to be of Elsterian age (0–2.5 m);
10. Reddish Middle-Devonian (Aruküla Fm.) sandstone.

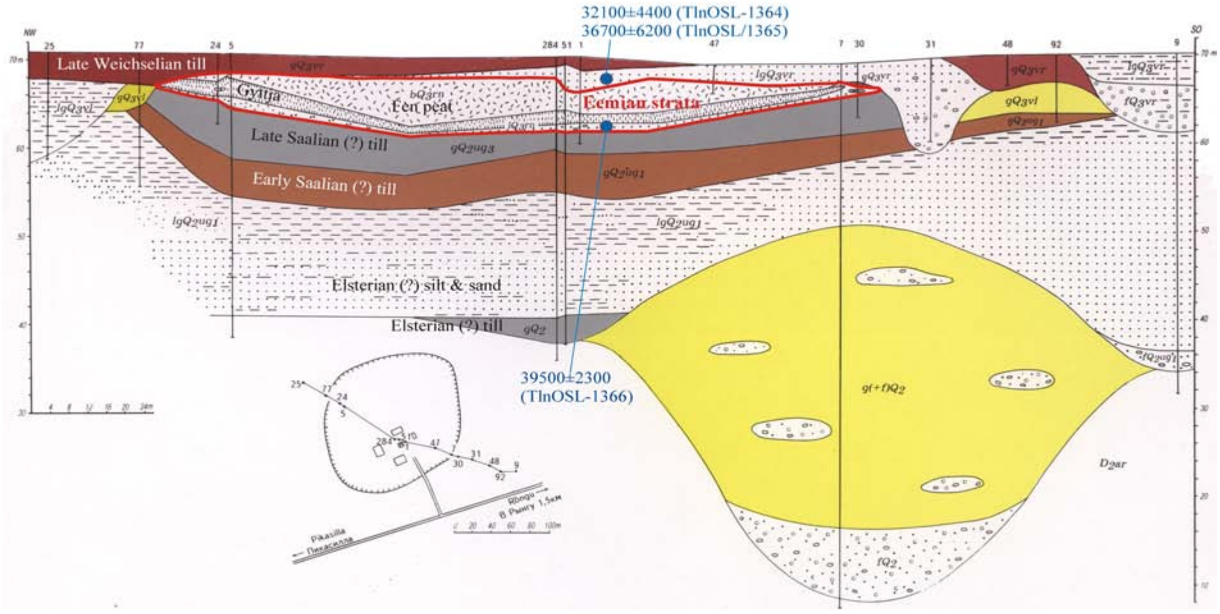

Fig. 3. Cross section at the Rõngu site, modified from Kajak 1995, OSL ages from Kalm 2006

E.Liivrand (1984, 1991) identified Eemian pollen assemblages (E1-E7: E1 = *Betula*; E2 = *Betula* & *Pinus*; E3 = *Pinus*, *Betula*, *Querqus* & *Ulmus*; E4 = *Querqus* & *Ulmus*; E5 = *Tilia*; E6 = *Carpinus*; E7 = *Picea*) at the section (Fig. 4). Also two TL dates for tills under (≥110 000 TL yr) and above (≥75 700 TL yrs) of the interglacial deposits are available from the late 1970-is (Kajak et al. 1981), the time of an over-optimistic approach to the developing TL method. Although the dating of tills itself, because of their highly variable genesis and problematic zeroing, is very questionable, the obtained TL age of the till (≥75 700 TL yrs) above the interglacial deposits raised a question about the age of the very thin (±2 m) surface till, that was considered to be of Late Weichselian age.

In 1999, the Rõngu section was re-examined and the fine-grained laminated fluvial sand above and below the interglacial strata were dated using the OSL method. The dates obtained (32.1 ± 4.4 from above, and 36.7 ± 6.2 and 39.5 ± 2.3 ka BP from below; Kalm 2005, 2006) seemingly support the ages (≥ 30 ka BP) of the earlier ^{14}C dates (Liiva et al. 1966) of organic

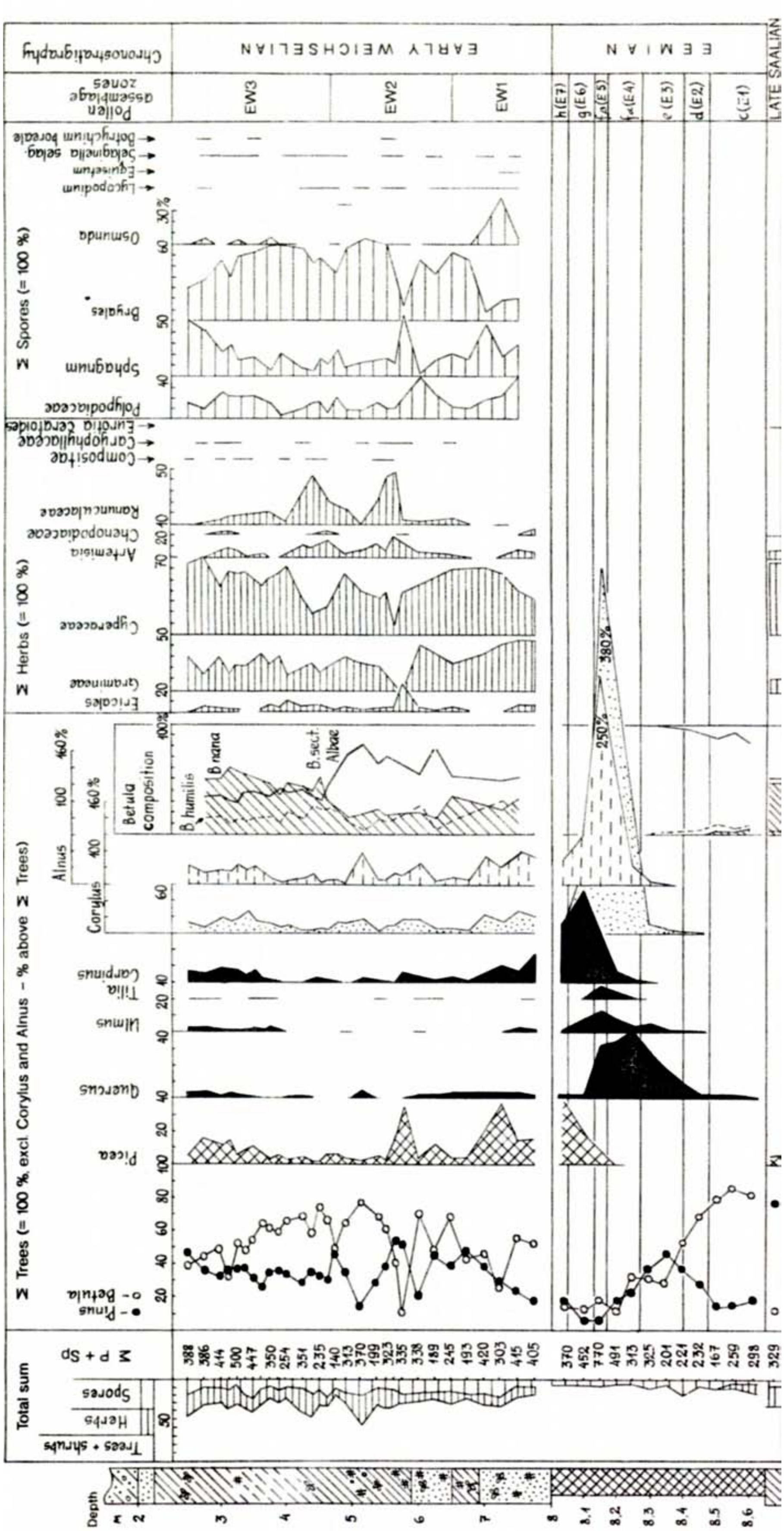

Fig. 4. Pollen diagram from the deposits of borehole No. 2 at the Rõngu site (Liivrand 1991).

sediments. However, palynological investigations have confirmed an Eemian age of these deposits and their correlation to other Eemian sections (Liivrand 1984, 1991). Pollen assemblages indicate that the entire Eemian vegetation cycle is present at Rõngu (Liivrand 1984, 1991). Assuming that the pollen analysis and bedding conditions in Rõngu site are correct, the OSL ages of deposits containing Eemian erratics reflect glaciotectonic dislocation, in which Eemian deposits have been placed into the Middle Weichselian sands. The Late Weichselian date for the dislocation is supported by the fact that sands with Eemian erratics are covered by only one thin bed of typical to southern Estonia reddish-brown Late Weichselian till.

Since the age of the interglacial deposits at Rõngu is not firmly established and new advanced dating methods have recently become available, we have returned to the long-studied site once again. In 2009, we drilled the site one more time and took core samples of peat and gyttja for U-series dating, and silt for OSL dating.

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STOP 3:

OTEPÄÄ UPLAND – AN INSULAR-LIKE UPLAND WITH HUMMOCKY TOPOGRAPHY

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Insular heights

The terrain in the SE sector of the Scandinavian glaciation is characterised by two main types of macro-topography: hummocky island-like uplands and glacial lobe depressions with almost flat or slightly undulating relief. The island-like uplands or so-called insular heights are spread widely and regularly in the NE part of the East European Plain.

The origin of the insular heights has earlier been described as ice marginal formations (end moraines) or disintegration features, formed in dead ice conditions. I.Danilans (1965) believed that randomly oriented hummocks on insular heights are inevitably dead-ice formations. Serebryanny and Raukas (1966) interpreted the heights as formations occurring in ice-divide zones, and Basalykas (1969) attributed their genesis to the concentration of glacial deposits around the bedrock cores.

Fig. 1. Glacial insular heights get their names due to their island-like position in the topography where they raise high above the surrounding lowlands. In Estonia Otepää and Haanja Heights are the typical accumulative insular heights and Pandivere and Sakala Heights have a bedrock core but very thin Quaternary cover being thus known as bedrock insular heights.

Due to their insular position in the topography these uplands raise relatively high above the surrounding lowlands (Fig. 1), reaching the altitude of 330 m, and in relative height up to 220 m. They occur in submeridional belts of 2–4 such forms (Fig. 2) and each of them with an area of up to 10.000 km^2. The heights have a bedrock core and they are composed of number of till beds and related intermorainic deposits. The normal bedding of deposits is highly disturbed, both horizontally and vertically, with considerable up-thrusting and folding, which complicates stratigraphical correlations. The markedly thick Quaternary cover on the heights (up to 300 m) consists of deposits from several glaciations, while the till and glacioaquatic deposits from the last glaciation are most significantly reflected in the glacial morphology.

Surface morphology displays the larger landforms at the higher positions and smaller on lower altitudes. At the higher altitudes, the hills are normally made up of thrust-type basal till, penetrated by ice-pushed fold structures and overlain by proglacial clays. Also specific elevations, dome-shaped hillocks (20–50 m), consisting of dislocated glaciofluvial deposits, are attributed to the central higher parts of the insular heights. The second level of the glacial

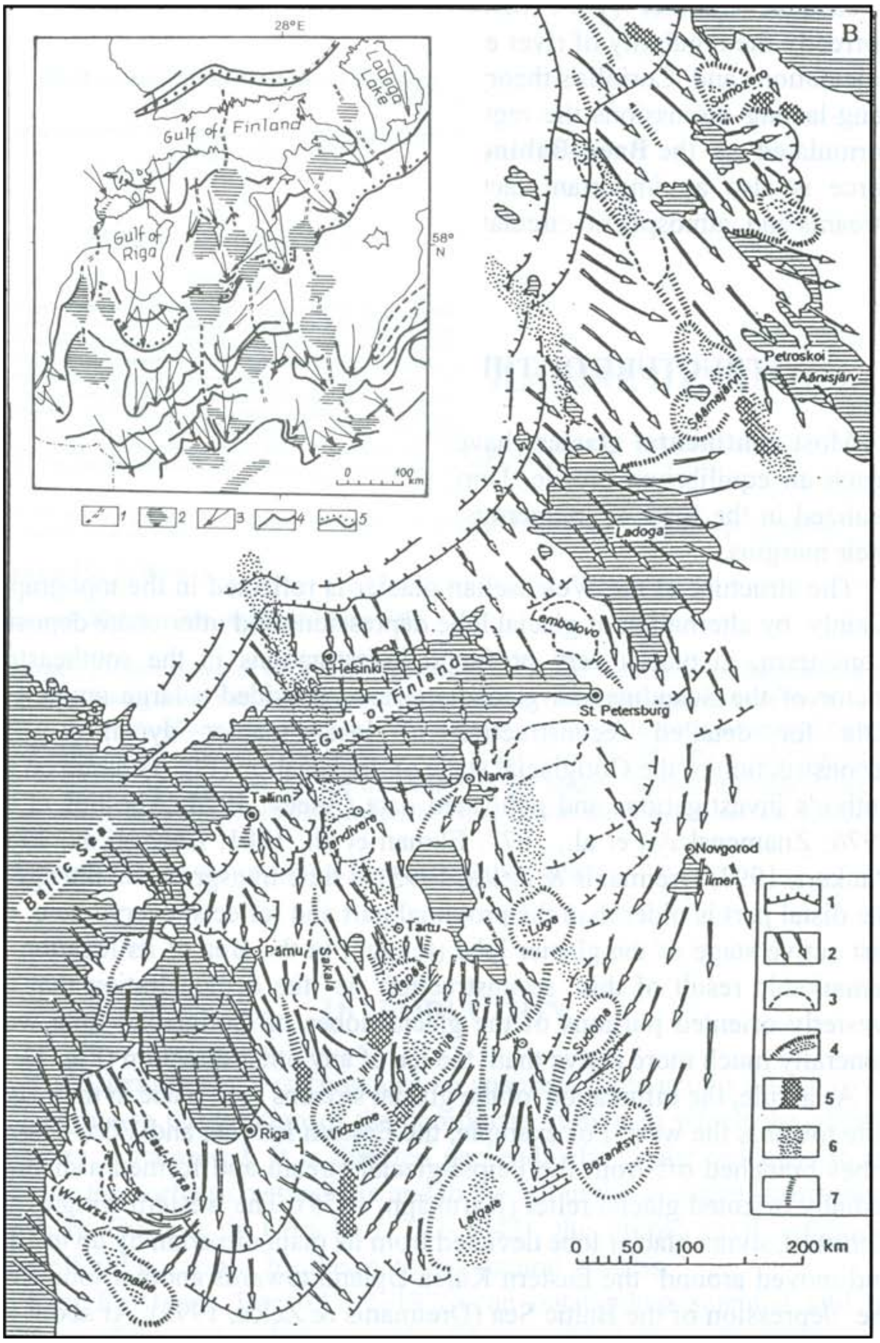

Fig. 2. Glacial dynamics and morphogenesis in the southeastern sector of the Scandinavian glaciation (compiled by R.Karukäpp using published data). 1–glacier margin; 2–ice-flow direction at the final stage of its activity; 3–bedrock upland at the ice-divide; 4–accumulative insular heights; 5–interlobate complex of landforms; 6–ice-divide between the glacier flows; 7–local ice-divide.

features around the highest central parts of the insular heights is represented by combination of a medium-sized hillocks (10–25 m) and small mounds (<10 m). The peripheral part of heights consists of small hummocks and marginal formations.

Complicated history of development of insular heights – a complex meso-relief forms, is explained through subglacial, englacial, marginal and stagnant ice stages. During all glaciations, the formation of insular heights started in ice-divide zones with subglacial accumulation of till around the bedrock elevations. As the deposition was more intensive at the proximal slopes, the present heights are located somewhat north or northwest of the bedrock elevations. Obstacles in the subglacial topography and sediment load led to the stagnation of thick glacial ice in the central area of present day heights. During the englacial stage, the glacial sole rose and stress increased in the basal part of the glacier. Folds and overthrust were formed in subglacial deposits. Probably the area of stagnation gradually rose and rapid accumulation took place in widening transition area between stagnant and active ice. This stage was followed by evident marginal accumulation during the ice recession. Small end moraines and the hilly topography were formed at the peripheral zone of the heights together with proglacial sedimentation. During the final stage, the dead ice designed the topography of the heights. Melting of dead ice blocks, glaciokarst, slope and erosional processes were prevailing. The timing of final decay of dead ice is not clear. In Estonia alone more than 20 sites are known where organic layers in insular heights are known in the upper till or between two till layers. Diversity of dates including several clearly too young ones, makes the correlations complicated and at least partly attributes the buried organic layers to glaciokarst or younger slope processes. Paludification, erosion, overwash processes and the accumulation of lacustrine and alluvial deposits were the most important landscape forming processes during the Holocene.

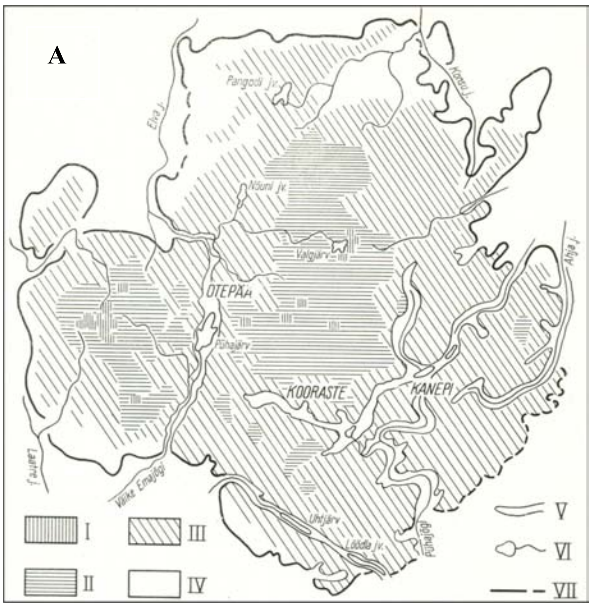

Fig. 3. Topography of the Otepää Upland, SE Estonia. I–heights above the altidude of 200 m; II–15-200 m a.s.l.; III–100-150 m a.s.l.; IV–<100 m a.s.l.; V–deeper valleys; VI–rivers and lakes; VII–foot of the upland.

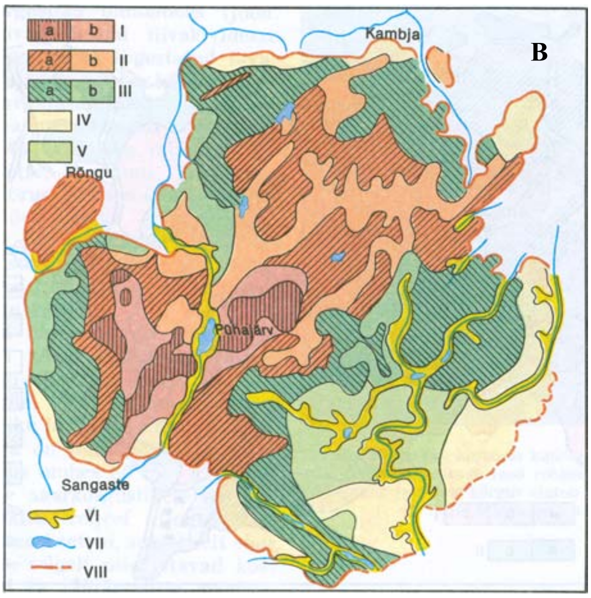

Fig. 4. Distribution of relief types in Otepää Upland. I–II–hummocks composed of till or till covered glacioaquatic deposits; III–hummocks composed glacioaquatic deposits (glaciofluvial and glaciolacustrine kames; IV–till or glacio-lacustrine abrasional plains; V–outwash plains; VI–valleys; VII–rivers and lakes; VIII–foot of the upland.

The Otepää Upland

The Otepää Upland in southeastern Estonia has a bedrock core, thick Quaternary cover with complicated structure, hummocky topography and clear slopes. The total area of the Otepää Upland is 1180 km^2; with its longest range both, from north to south and east to west reaching 40 km. The average altitude of the foot of upland is 88 m. 19 higher points of the uplands topography rise over 200 m, the highest point, Kuutse Hill reaching the altitude of 217 m (Fig. 3). A depression, holding the Lake Pühajärv and trending from north to south, divides the Otepää Upland into two parts. In the central part large hummocks with the relative height of 50–60 m prevail while in the peripheral part of the upland small hummocks and elongated hills as well as abrasional till and glaciolacustrine plains together with outwash and glaciolacustrine accumulative plains are present (Fig. 4). In SE part of the upland 30–40 m deep valleys are abundant.

The distinction between large, medium scaled and small forms is made on the basis of their relative height, which is over 25 m, 10–25 and up to 10 m respectively. The diversity of relief forms (Fig. 5) include till formations, moulded by the glacier, end moraines, kames and various other forms. The dominating landforms are kames of different type. In addition, the distribution of relief forms of different origin demonstrates the correlation with the altitude. The highest level of glacial accumulation at the Otepää Upland is characterized by the landforms with a top of glaciolacustrine deposits among the landforms with complicated glacially deformed structures. Middle level of glacial accumulation constitutes the distribution area of till hummocks or till covered kames together with glaciofluvial and glaciolacustrine kames. In addition, push type end moraines can be found. The lowest level of glacial accumulation is characterized by the proglacial sedimentary environment reflected in undulating glaciolacustrine plains in the outlying areas of the height while the water level changes in proglacial lakes are reflected in numerous terraces at the slopes of deep valleys. Such a distribution of landforms of different origin points to the beginning of land forming processes in the more elevated parts of the upland. Those areas became ice-sheds where the dead ice was formed. It is believed that at the same time in lower parts of the upland as well as in the surrounding areas the ice was still active. Here the stagnation of ice and formation of dead ice took place much later. Proglacial bodies of water were formed after the ice margin finally retreated from the area. Relief analysis and fabric orientation have been used in reconstruction of the ice flow directions (Fig. 6).

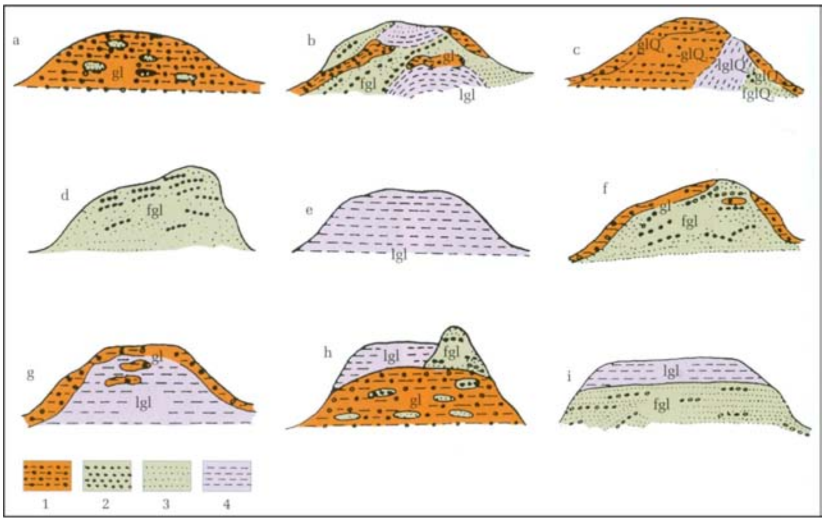

Fig. 5. Different landforms characteristic to the Otepää Upland (after Kajak, 1963). Aa–till hummock; b,c–push-moreine forms; d–glaciofluvial kame; e–glaciolacustrine kame; f–till covered glaciofluvial kame; g–till covered glaciolacustrine kame; h–complex forms composed of till and glacioaquatic deposits; i–platoo-like hill composed of glacioaquatic deposits. 1–till; 2–pebbles; 3–sand and gravel; 4–laminated sandy-clays.

Hummocky topography, mosaic pattern of sediments/soils, weak infiltration qualities and intensive erosion turn the Otepää Upland inconvenient for agricultural purposes. Thus, this hummocky area was settled much later than the lowland areas. However, the picturesque landscapes, abundance of lakes and long winters (snow cover at the uplands lasts at least two weeks longer than in lowland areas) provide the upland with notable recreational and sports value. Surroundings of Otepää serve as an international cross-country skiing centre, which has given also two aboriginal Estonian Olympic champions!

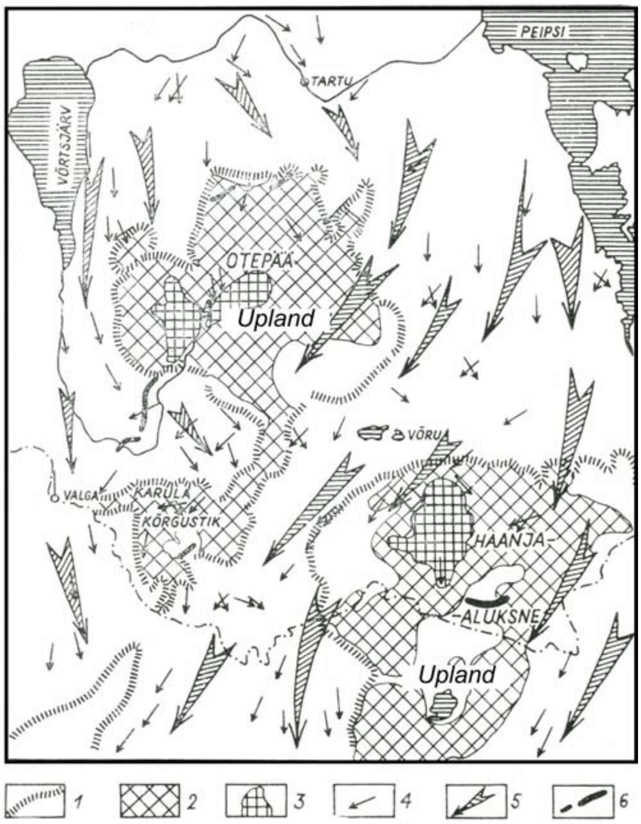

Fig. 6. The formation of the Otepää and Haanja insular heights and the ice-lobe dynamics in the end of the Late-Weichselian glaciation. During the englacial and peripheral glacial accumulation stages in the development of insular heights, ice lobes in the surrounding lowlands were still active. 1–foot of the height; 2–small and medium-size hummocky topography; 3–hilly topography with platoo-like kames; 4–orientation of the elongated clasts in the till; 5–direction of the proposed ice-movement; 6–end moraines.

STOP 4:

SECONDARY CARBONATE PRECIPITATE IN OUTWASH DEPOSITS, MATSI SAND PIT, SE ESTONIA

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The Matsi pit (57°52′41′′N, 26°16′36′′E) is located on the top of an elongated and flat-topped end moraine ridge (Fig. 1) which highest point (72.2 m a.s.l.) is excavated. The ridge has a relative height of about 15–20 m and an asymmetrical cross profile with gentle down-glacier distal (eastern) slope, and a higher and steeper up-glacier proximal (western) slope. About 800 m long and 300 m wide N–S stretching ridge, belonging to Sangaste-Laatre-Valga end moraine zone, has formed in front of the local Võrtsjärve ice-lobe during the Otepää stade at about 14.5 ka cal ^{14}C yrs BP.

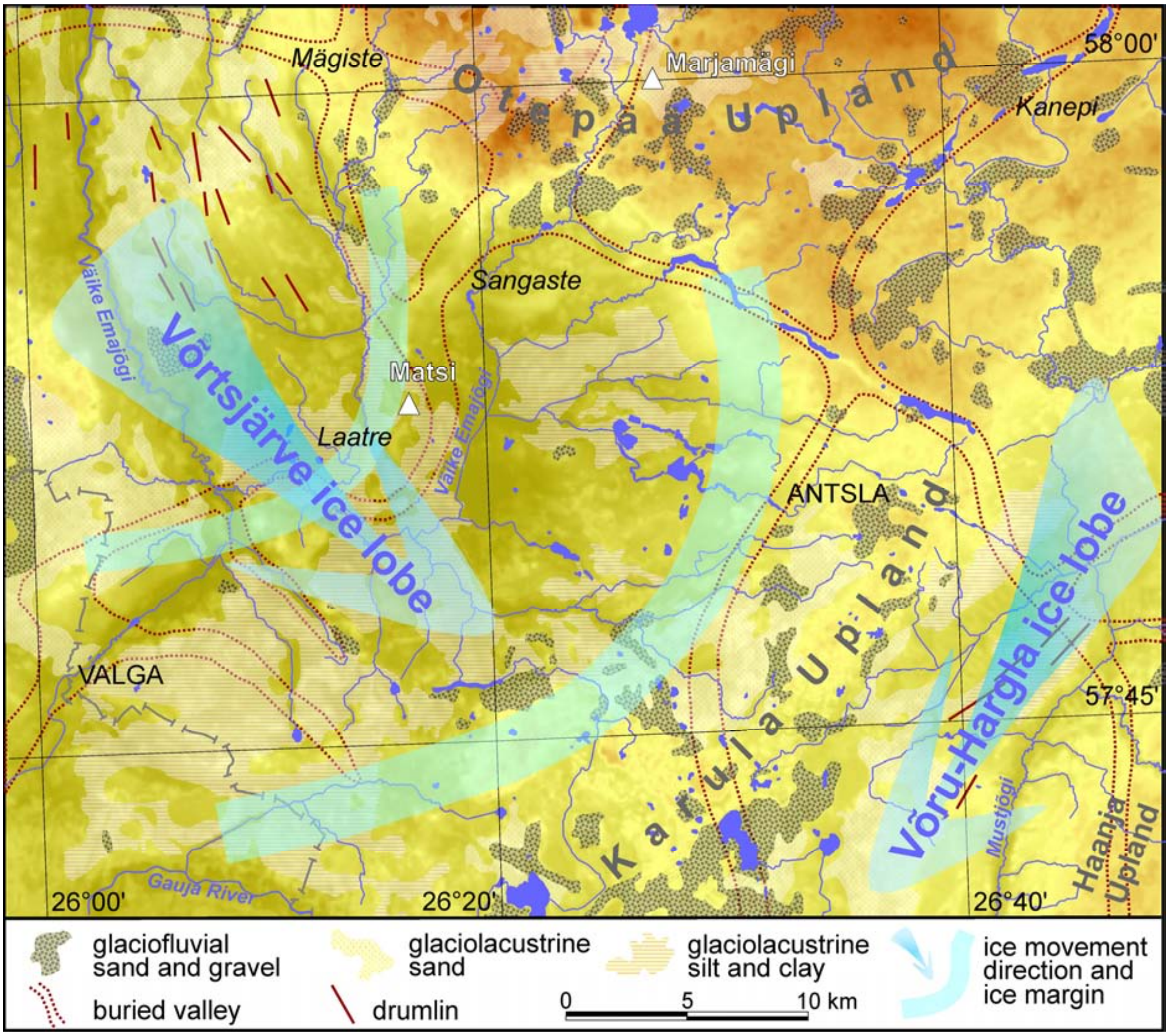

Fig. 1. Topography, geology and ice flow pattern in surroundings of the Matsi site.

The overall thickness of the Quaternary deposits reaches up to 40 m in the end moraine ridge at Matsi. However, in the neighbouring pre-Quaternary valley, the thickness of the Quaternary deposits reaches 159 m at Mägiste and 75 m at Sangaste and Väike-Emajõgi River valley (Fig. 1).

The end moraine ridge is composed of three sedimentological units:

1. The lowermost unit is compact and massive greyish-brown matrix-supported sandy till, deposited subglacially as lodgement till, typical for southern Estonia.
2. The bottom till is overlain by sorted fine to medium-grained sand (Fig. 2). The sand has horizontal to low angle (dipping to the east and southeast) planar cross-lamination. Sand was deposited in outwash fan during a standstill of the ice margin in its retreat from the marginal zone.

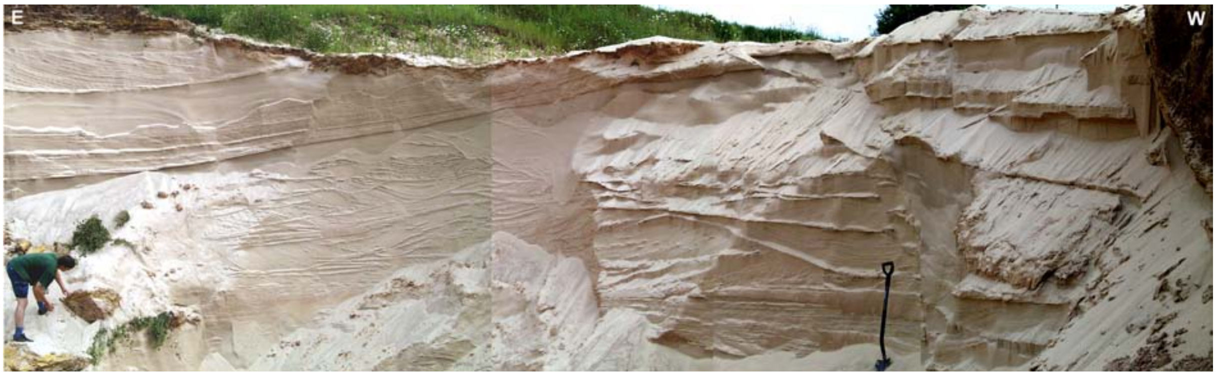

Fig. 2. Fine-grained, cross- and ripple-bedded sand facies in Matsi pit.

3. 3 m thick, massive to laminated sandy-silty till of the last glaciation covers the sand unit discordantly. The till bed has a sharp and planar lower boundary, indicating that no diffusive mixing with the sediments below has happened, or it was restricted to a very thin mm-scale zone only. Highly attenuated, folded and boudinaged sand lamina in the till represent separate sequential phases of accumulation of deformation till, most likely during a short-lived ice advances.

Fig. 3. Cemented piles in fine-grained sand.

Secondary carbonate precipitates in outwash sediments

Carbonate precipitation in outwash deposits has taken place beneath the upper till layer. Cemented aggregates are laterally restricted to few meters but form up to 3 m high vertical piles (Fig. 3). Carbonate precipitate is intruded into the till above only for few cm. Cemented piles of sand occur above the modern water table.

The carbonate cement is common in matrix-rich sandy facies, filling almost all inter-granular pore space (Fig. 4). Micromorphologically the matrix-supported cement is present mostly as micrite (<4 μm) and microsparite (4–10 μm) with angular equant to rhombohedral calcite crystals. Sparitic (up to 50 μm) calcite assemblages fill larger inter-granular spaces or microfractures. Micrite is usually located close to grain surfaces, indicating that precipitation of micrite preceded that of microsparite and/or sparite. Multilayered cement and dissolution traces on the surface of calcite crystals suggest, that episodes of cement precipitation alternated with periods of dissolution. The cement is expected to have formed in conditions of water-saturated sediments.

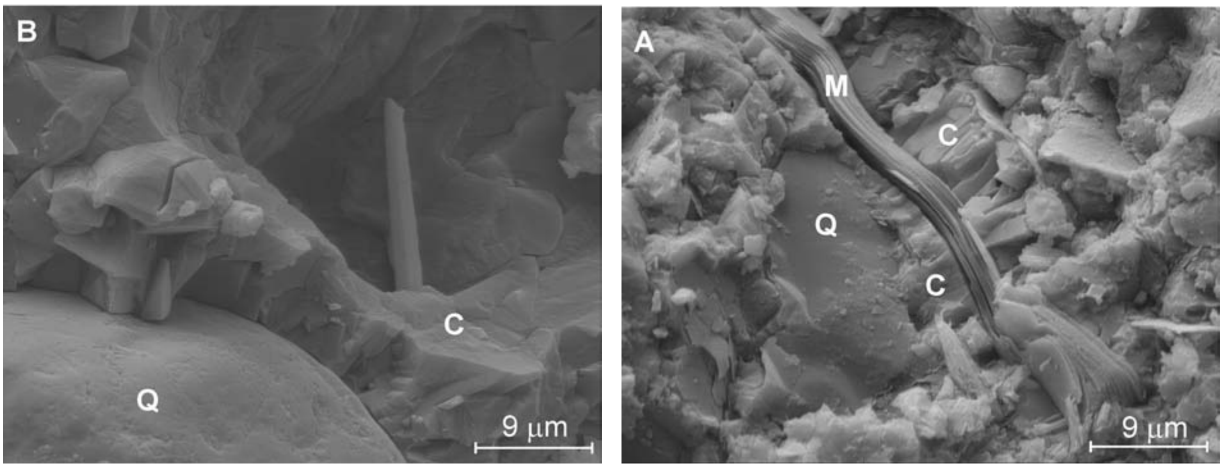

Fig. 4. SEM images of the calcite cement: A – in till matrix, B – in sand matrix. Minerals: Q – quartz, C – calcite, M – mica.

We believe that the secondary calcite at Matsi site may have precipitated in proglacial conditions when the ice margin was close to the site. Calcite precipitation occurred in conditions of enhanced groundwater pressure gradients and in presence of partly frozen sediments or permafrost. The upper till layer acted as a barrier to the water flow, diverting groundwater circulation in outwash deposits. Isotopic composition ($\delta^{18}O$ –7.7…–8.9‰ VPDB; $\delta^{13}C$ –8.6…–10.2‰ VPDB) of secondary calcite shows, that the solute-bearing waters were enriched in light carbon, and probably somewhat mixed with meteoric and surface waters.

STOP 5:

NEW LATE GLACIAL CHRONOLOGY, ENVIRONMENTAL AND CLIMATIC CONDITIONS IN SOUTHERN ESTONIA: EVIDENCE FROM LAKE NAKRI

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The synchronisation of palaeoenvironmental events in the North Atlantic region during the last Termination has undergone tremendous steps in the last years (Rasmussen et al. 2006, Lowe et al. 2008). Keeping this in mind, it is striking, that although there are about 100 age estimations from the Estonian late glacial sediment sequences (Kalm 2006), none of them can be associated with biostratigraphical studies. This means that practically all of the lateglacial sequences in Estonia, and eastern Baltic area in general, are not independently dated and are described solely by comparison of pollen etc. zones and pollen percentage data, although already since M.Davis (1964) the importance of pollen accumulation rates (PAR) in the late glacial was shown. Moreover, there are no ^{14}C-dated investigations on lateglacial vegetation succession in latitudinally adjacent areas, except some sites in eastern Karelia, Latvia and in Lithuania.

The excursion stop at Nakri locality discusses a detailed well-dated late glacial floristic re-colonisation and vegetation succession in southern Estonia between 14 000–9000 cal. ^{14}C yr BP, as revealed by palynological, macrofossil, loss-on-ignition and ^{14}C data.

The new investigated lateglacial site, Lake Nakri (0.9 ha, 48.5 m a.s.l. 57º53.703 N, 26º16.389 E) is situated in southern Estonia and is today a small, hard-water, shallow (3.2 m deep) water-body, surrounded by fen and mixed conifer forest. The lake basin with a catchment area less than 0.25 km^2 lies within the Otepää ice marginal zone, which formed approximately 14 700–14 500 cal. ^{14}C yr BP.

Terrestrial macrofossils (preferably small branches, *Dryas* leaves and *Betula* catkin scales) identified in the course of macrofossil analysis were ^{14}C dated by AMS method. The chronology of the Nakri lake sediment sequence is based on weighted average calibration of AMS radiocarbon dates (0=AD 1950), based on the INTCAL dataset, from terrestrial plant remains of nine levels, fitted into the OxCal 4.1 deposition model at 2σ confidence level.

The investigated late glacial – early Holocene sediments of Lake Nakri were divided into six sedimentological units, of which the lowermost unit 1 (1656–1500 cm below the sediment surface) was beige silty clay, getting increasingly sandy towards the bottom, possibly varved clay unit 2 (1500–1464 cm) was black-coloured silt with up to 7% of organic matter. Unit 3 (1464–1262 cm) was organic-poor beige clayey silt to silty clay, which became more organic-rich towards the upper boundary, and changed into black coarse detritus gyttja of unit 4 (1262–1240 cm). Unit 5 (1240–1207 cm) was distinctly laminated gyttja (possibly annually laminated), which gradually turned into homogeneous brown gyttja of unit 6 (1207–320 cm; i.e. top of sediment).

The Estonian late glacial vegetation change, in general, documents Older Dryas arctic conditions, an Allerød warming with up to 80% of pine pollen, a Younger Dryas herb-dominated cooling, and the following Holocene warming. Up to now no indication of a Bølling warming has been recognised, nor are there absolute chronologies to support the vegetation evidence, thus the terms such as Allerød and Older Dryas were more or less of a local PAZ's.

The deposition model inferred that the area around Lake Nakri, immediately south of the Otepää ice marginal zone, deglaciated just before 14 000 cal. yr BP, i.e. since the onset of the Greenland event GI-1c. Combined corroborating evidence from macrofossil and pollen analysis showed a *Betula nana – Salix herbacea – Dryas octopetala – Saxifraga* sp. – *Empetrum nigrum – Artemisia – Helianthemum nummularium – Chenopodiaceae* dominated shrub and herb tundra community with *Juncus* on wet ground at 14 050–13 400 cal. ^{14}C yr BP. Although the initial development of flora and landscape around Nakri basin corresponded with the relatively warm phase of GI-1c on Greenland, the sediments were highly minerogenic, PAR was very low and pollen highly corroded. Terefore it is presumed that up tp 40–80% of birch pollen and 100% of alder and hazel pollen was inwashed from earlier sediments, preferably from the Eemian flora, incorporated into the Late Weichselian till, surrounding the lake. PAR values of both *Betula* and *Pinus*, that are under 500 grains cm^{-2} yr^{-1} in the Lake Nakri at 14 050–13 400 cal. ^{14}C yr BP clearly show treeless conditions in the area, which is supported by lack of tree-type macro-remains in the sediments, and agrees with the general idea of late glacial evolving primary vegetation.

LOI of the sediment formed between 13 400 and 12 850 cal. ^{14}C yr BP raised from 2% to up to 7.5% and *Betula* PARs reached over 4000 grains cm^{-2} yr^{-1}. The share of corroded *Betula* pollen grains is low, confirming that most of the birch pollen is local. Moreover, finds of *Betula* sect. *Albae* seeds confirm the presence of tree birch in the area. As to pine, its PAR surpassed the local presence limit of 500 grains cm^{-2} yr^{-1} between 13 300–12 850 cal. ^{14}C yr BP and the forest limit of 1500 grains cm^{-2} yr^{-1} between 13 150–12 850 cal. ^{14}C yr BP. So far there was no evidence of late glacial pine macrofossils in Estonia, until a find of a single *Pinus stomata* at 13 300 cal. ^{14}C yr BP at Nakri. Inside the ~500-year long warming, associated with the GI-1a and GI-1c warm events a short cooling can be detected from a drop in birch and pine PAR values, centred to 13 100 cal. ^{14}C yr BP and connected with the GI-1b cooling, which in western European records has been recognised as the Gerzensee fluctuation.

At 12 850 cal. ^{14}C yr BP since the start of the Younger Dryas (GS-1) cooling the vegetation was again dominated by herbs and dwarf shrubs and the corrosion index indicated higher input of inwashed pollen and lower local tree pollen production. *Artemisia – Dryas – Helianthemum– Juniperus – Betula nana* dominated in the pollen record whereas the macrofossil evidence showed high concentration of dwarf birch and *Dryas octopetala* while tree birch and tree macroremains altogether were missing again. Gyttja sedimentation started gradually around 11 650 cal. ^{14}C yr BP at the late glacial – Holocene boundary, and within sediments of 300 years LOI content reached over 25%. The Preboreal oscillation around 11 400 cal. ^{14}C yr BP is clearly seen in Lake Nakri as a climatic reversal on the background of general Holocene warming, as a sharp decrease of LOI content, and decline in PARs of pine and birch and higher percentages of *Juniperus* pollen. The vegetation response at the Younger Dryas – Holocene boundary after the PBO was rapid forestation and loss of the late glacial floristic component.

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STOP 6:

HOLOCENE LAKE RESPONSES TO LAND-USE INDUCED CATCHMENT CHANGES AND DISTURBANCES IN SOUTHERN ESTONIA: LINKING LAND AND LAKE

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Our current knowledge shows the ability of biostratigraphical methods to pinpoint traces of human impact on environment already during the Mesolithic time in Estonia (Poska 2001). Recent palynological observation by Poska & Saarse (2006) that *Triticum* type pollen accompanied by *Cannabis* type (hemp) pollen is recorded at ca. 5600 BC at Akali, middle eastern Estonia, can be interpreted as possibly the first traces of the acquaintance of foragers with crop farming through contacts with southern European agrarian tribes already in the Late Mesolithic. During the Bronze Age ca. 2000 BC, the clear transition from hunting-gathering based economy to crop cultivation farming economy took place in the northern Estonian coastal alvar areas with rich soils suitable for early rural practices (Poska et al. 2004). This transition led to significant human impact on the local environment, brought considerable change to structure of landscape and composition of the land-cover and induced the deforestation of many areas in Estonia.

So far the biostratigraphical studies in Estonia have been generally aimed to investigate the past impact of land use on terrestrial ecosystems. Much less is known how did early human land-use influence aquatic ecosystems. Therefore we initiated a project where we plan to utilise a variety of paleoecological proxies to investigate lake response to changing agrarian activities through the whole period of human impact on the southern Estonian landscape, i.e. the last 6000 years. Pollen evidence allows to track prehistoric human impact on vegetation in the lake catchments area since the introduction of farming, whereas diatoms preserved in sediments provide precise data for reconstructing past trophic changes in lakes.

Lake Tollari, a small strongly stratified hard-water eutrophic lake (57º45´08´´N; 26º20´27´´E, max. depth 10.5 m, surface area 5.7 ha), is positioned on the northern slope of the Karula Heights within a landscape that probably has been heavily impacted by human rural activities over the millennial time-scale. The recovered sediment core consisted of 10.5 m of Holocene gyttja underlain by 1.5 m of lateglacial silts. Chronology for Lake Tollari sediment sequence was developed by 13 radiocarbon dates. Sampling interval for biostratigraphic measurements is after every 40–60 years for the time period of 4000 BC – AD 2000 and after every 100–150 years for the time period 9700 – 4000 BC. The diatom analyses are completed, however the pollen analyses are still in progress. The diatom composition data and diatom inferred lake surface water total phosphorus concentrations suggest that influential change occurred in the lake water quality around 1200 BC, in-lake nutrient concentrations increased and the lake switched from mesotrophic state to eutrophic condition possibly in connection with external man-made catchment forcing.

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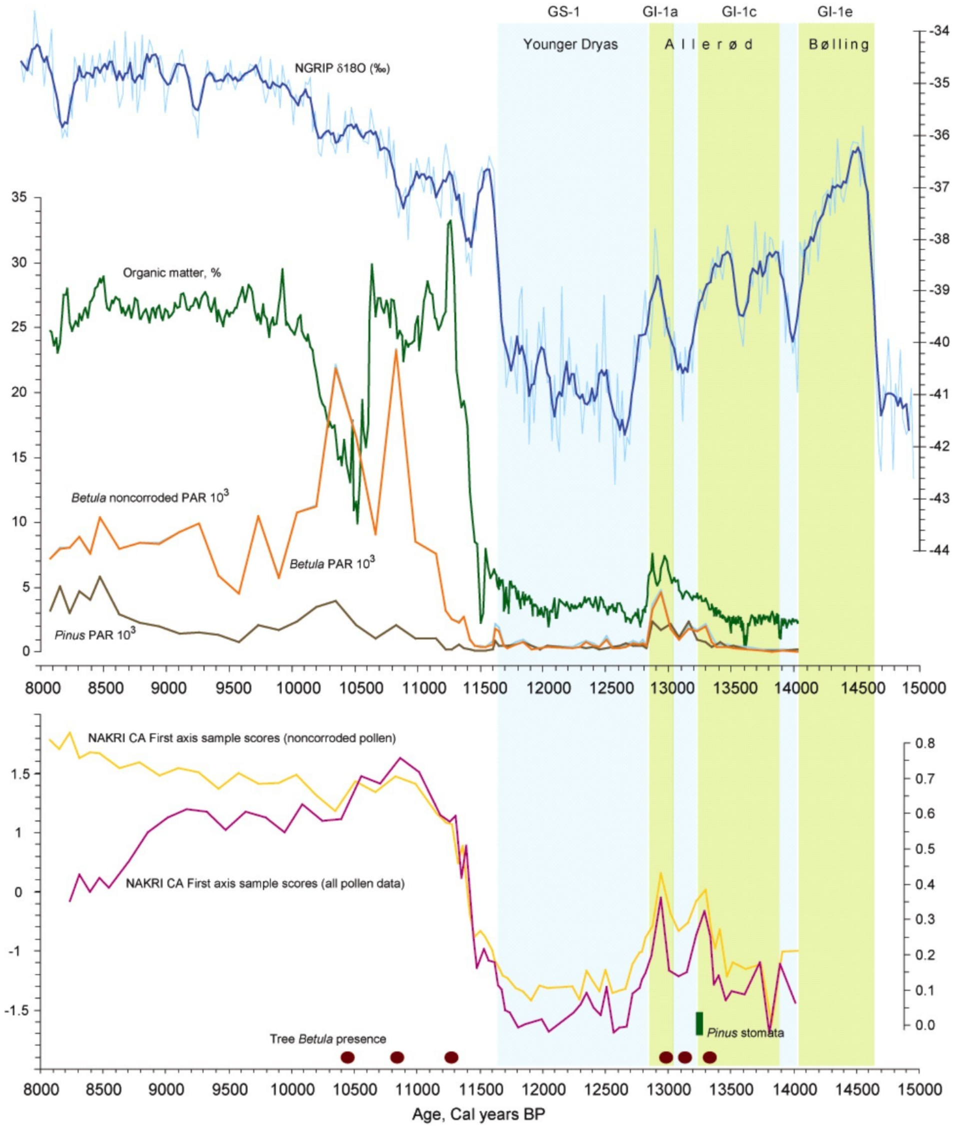

STOP 7:

ALTERED ICE-MARGINAL FORMATIONS OF HAANJA (LINKUVA, NORTH-LITHUANIA) STAGE AT DORES, NORTH VIDZEME LOWLAND, LATVIA

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Ice-marginal formations at the farmhouse Dores is a constituent part of the complex of the ice-contact and proglacial deposits associated with the recessional moraine. The recessional moraine itself stretches for 7 km in NNW-SSE direction from Valka (Fig. 1). It lies at the proximal side of the outermost endmoraine that delimits the position of the Burtnieks (North Vidzeme) ice lobe at its maximal extent during the Haanja (North Lituanian, Linkuva) stade (ca. 13.500 BP according to Raukas et al. 2004). The southern continuation of the recessional formations is interrupted by the River Gauja valley, and altered to a great extent by subsequent meltwater and aeolian activity. The recessional moraine ridge is fragmented by erosional features into several segments. The highest part of the segment at the Dores site reaches the altitude of 82.8 m, being about 900 m long, 300 m wide and up to 15 m high with asymmetrical cross profile. The proximal slope is much steeper than the distal one (13º and 5º respectively).

The test drilling data and detailed studies of the internal composition of the ridge in sections of the Dores sand and gravel pit reveals the following major lithostratigraphic "units":

1. The lower till were only reached by boreholes and was likely deposited as a result of subglacial deposition during the glacier advance. It consists of reddish brown sandy clay or clayey sand diamicton that is underlain by Middle Devonian sandstone, siltstone and clay. The lower till builds up the base of the ridge. The till surface is slightly undulated (relative height less than 10m). These buried topographic unconformities supposedly resemble a drumlinized till superimposed by a transverse moraine ridge.
2. Sandy silt and silty sand of various thicknesses overlay the lower till. Drill records indicate that these glaciolacustrine sediments with maximum thickness up to 4.5 m are mostly present in the depressions of the till surface. These sediments have probably been eroded by meltwater streams from the higher altitudes.
3. Up to 6–10 m thick layer of stratified coarse gravel, gravely sand and various grained sand with interbedded silt overlies the lower till unit. The layer is thinning out in the eastern, i.e. distal direction. Besides distal sediments are finer-grained. Therefore this unit most likely represents the apex and distal sediments of a proglacial subaquatic fan.
4. Fine sand with interbeds of silt and coarser-grained sediments overlie the mentioned above stratified material. The thickness of these sediments varies from 1.5–2 m diminishing at the crest of the ridge. Upper part of the layer is disturbed by minor folds, faults, clastic dykes, cast and water escape structures. Its lower contact with glaciofluvial sediments is sharp and well-defined. The upper contact with the overlying diamicton differs from place to place being partly hard to define.
5. The upper till unit is brown clayey-sandy diamicton that contains sand and silt lenses. It is on an average 0.2–0.3 m thick and is recognized only on the upglacier slope of the ridge. Occasionally, the material from units 3 and 4 has been incorporated into the diamicton as a result of the shearing and thrusting. The diamicton could be interpreted as a flow till. However in some places the material becomes massive and resembles typical structure of a melt-out till

6. Glaciolacustrine sediments rest conformable on the upper diamicton forming up to 1 m thick layer on the western (proximal) slope of the ridge. The glaciolacustrine unit consists of varved silt and is very rich in dropstones. The upper contact with the uppermost glaciofluvial sediments is sharp and easily defined.
7. The upper glaciofluvial sediment unit comprises layers of various grained sand with rare well-rounded pebbles with normally graded bedding and gravely sand. The thickness of these sediments is 0.7–0.8 m. This sequence of glaciofluvial sediments occurs only on the western slope but gravely-sand various grained sand with random pebbles are encountered also in other parts of the ridge.
8. The entire Pleistocene sequence is covered by up to 0.5 m thick layer of drift sand.

Based upon internal structure of the ice marginal ridge at Dores it could be concluded that most part of the material composing this landform was deposited in the ice dammed lake during a temporary standstill of the edge of the glacier. Glaciolacustrine sedimentation was interrupted by minor subsequent glacier readvance when the upper till unit have been deposited and together with the older glacioaqutic sediments pushed up creating the recessional moraine ridge. Following the ice recession from the area aeolian activity and paludification have been the most important agents in the formation of present topography.

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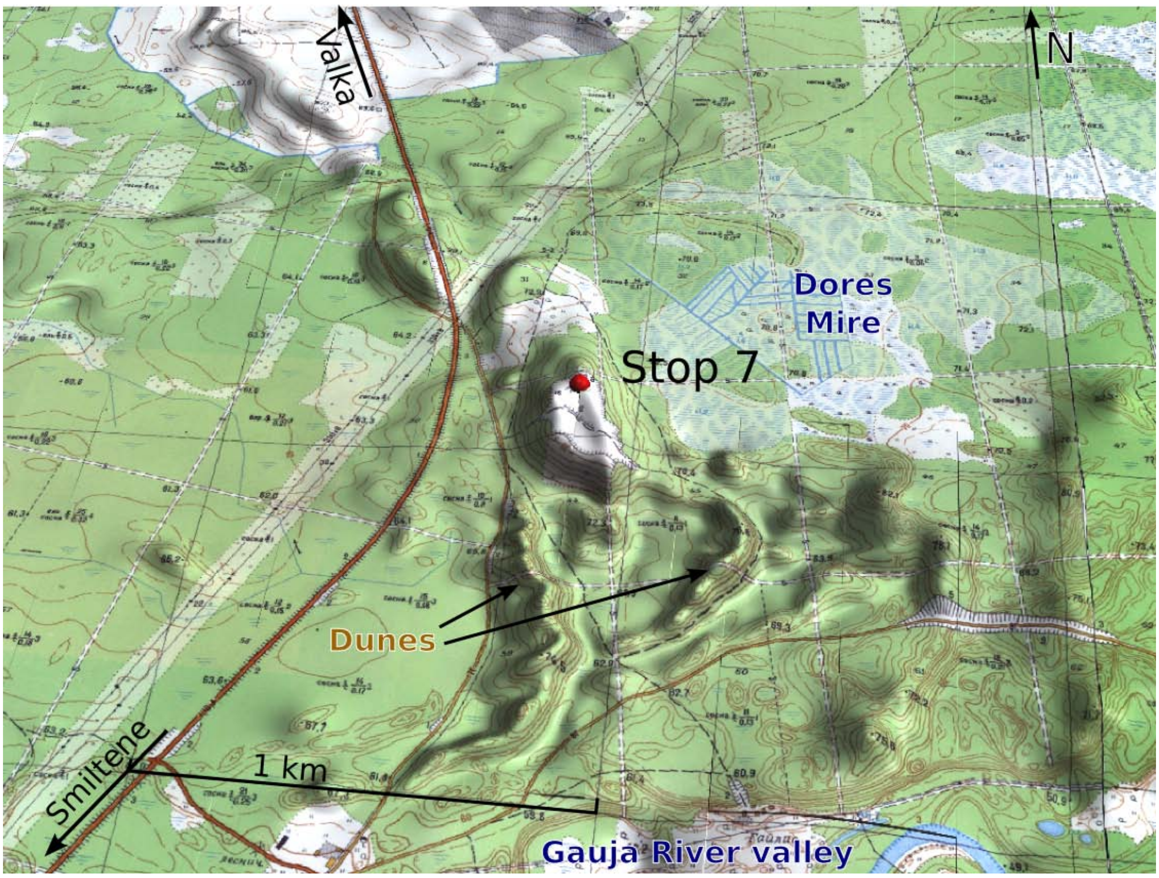

Fig. 1. Location and shaded topography of the Dores site.

STOP 8:

HISTORY OF THE DEVELOPMENT AND PALAEOGEOGRAPHY OF ICE-DAMMED LAKES AND INLAND DUNES AT SEDA SANDY PLAIN, NORTH WESTERN VIDZEME, LATVIA

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The Seda Plain occupies the hypsometrically lowest part of the Northern Vidzeme lowland. It lies between the Vidzeme Upland, Burtnieks drumlin field, Sakala Upland, Aumeisteri and Karula interlobate highs. The highest points (~100 m a.s.l.) and most pronounced topography relates to the inland dunes, while the lowest part is occupied by the valley of the River Gauja (30 m a.s.l.). History of the development of glacial lakes in the Seda Plain is very complex. Due to the undulated topography with alternating of the interlobate areas, local highs and depressions, and common inclination of the land surface of the glacial lowland towards the retreating glacier, melt waters could not drain freely, resulting in a series of ice-dammed lakes along the retreating margin of the Burtnieks ice lobe.

The plain lies in the local bedrock depression. The hypsometric position of bedrock surface, consisting of the Middle Devonian sandstone, aleurolite and clay, varies from 20 to 60 m a.s.l. The bedrock is covered with up to 5–7 m thick, presumably Late Weichselian till. This till overlies slate-blue clay of an uncertain age at the bottom of the buried valley near the Strenči town. In the most part of the Seda Plain the till is overlaid by up to 10 m thick layer of the glaciolacustrine fine grained sand or silty sand. The top of the glaciolacustrine sediments are reworked by aeolian processes over a vast area. Glaciolacustrine sand is the dominant sediment. The silty sand prevails only in the NE part of the plain. Up to a few meters thick clay bed was encountered between till and glaciolacustrine sand at the peripheral part of the plain, for example N of the River Gauja. At some locations course glaciofluvial sand and gravel underlay glaciolacustrine sediment. The deposition time of these glacioaquatic sediments is unknown.

During Late Glacial and at the beginning of the Holocene the plain was occupied by two large successive glacial metwater lakes – Smiltene and Strenči. The Smiltene ice-dammed lake formed between the slope of the Vidzeme upland and the stagnating Burtnieks (North Vidzeme) ice lobe. It occupied the southern and southeastern parts of the contemporaneous plain. The approximate area of the lake was ca 400 km^2.

The water level of this lake was changing gradually along with melting of the Burtnieks ice lobe. Therefore, ancient shorelines are morphologically underdeveloped. The most pronounced levels of the Smiltene proglacial lake can be fixed only by delta levels at 75 m a.s.l. (Brutuļi delta at Smiltene), 70 m and 65 m a.s.l. (Mustjegi (Mustjõgi) delta). Only some morphological evidence has been found for higher levels of this lake. Traces of the ice-dammed lakes being at the highest level have been also encountered in the northern part of the plain, for instance, the top of the Pentsils delta marks the existence of the local ice-dammed lake at the level of at least 80 m a.s.l. However, morphological and sedimentological evidence of the development of the meltwater basins at this level along the northern margin of the plain is sparse. That can be explained by small dimensions and short time of existence of such small meltwater bodies.

After drainage of the Smiltene ice dammed lake from 65 m to 60 m a.s.l. level, it was transformed into the Strenči proglacial lake. The shoreline at the level of 60 m a.s.l. is traceable at both ends (S and N) of the Seda plain. Thus it can be assumed, that during this phase the most part of the plain was ice-free. The shoreline of 52 m a.s.l. is better marked. According to

O. Āboltiņš (1971) the lower outlet of the River Gauja related to this level is situated downstream of the Strenči town. After the drainage of the proglacial lake to the level of 48 m a.s.l., only small remnant lakes were left in lower parts of the plain. Later most part of the area covered by these lakes turned into swamps and/or peat bogs. At its maximal level, the Strenči proglacial lake covers an area more than 800 sq. km.

The inland dunes are common for the most part of the Seda plain. The highest concentration of the dunes occurs in the central, northern, and northeastern part of the Seda plain as well as along the valley of the River Gauja. Absolute heights of the inland dunes vary from 46 m to 98 m a.s.l. with an average height of 65 m a.s.l. The maximal relative height of some dunes reaches 25 m, but on average, it is 4 m. On the basis morphological complexity of dune pattern, simple, compound and complex types of the inland dunes have been recognized. Simple dunes are rare, more common are compound and complex dunes, especially within dune complexes. It is possible to distinguish up to seven dune complexes.

In total 11 samples of dune sand were collected for OSL (SAR protocol) dating. Samples were processed at the Dating Laboratory of the Finnish Museum of Natural History in 2007 and 2008. Obtained results indicate that inland dunes forming sand were last bleached within timeframe from 6.4 ka to 11.9 ka BP. Most of the samples relate to the Preboreal, Boreal and Atlantic time (Fig. 1). The inland dunes with a topographically higher location of their base are younger than lower placed ones. Dating results reveal that formation of dune had been started in the Younger Dryas. It can be assumed that aeolian processes in the Seda plain territory began just after the disappearance of the Smiltene and Strenči proglacial lakes. It seems that the comparatively cold environment of the Younger Dryas was too harsh for continuous vegetation cover to form in the sandy parts of the Seda plain. Thus aeolian processes were ongoing in the most part of the territory. At Kauci (11.9±2.7 ka BP Hel-TL04142) and Silezers (11.8±2.4 ka BP Hel-TL04144) sites due to local factors (most likely difference in topography/soil moisture) the development of the inland dunes started to slow down already at the end of the Younger Dryas. At the beginning of the Preboreal time, temperatures started to rise and thus enhanced vegetation spread in the areas controlled by sandy aeolian processes, and thus stopped dune development. At first, only smaller, simple and separate dune formation was affected (sites Kampinas 10.2±1.9 ka BP Hel-TL04139, Dores 9.4±1.9 ka BP Hel-TL04140, Lushi 10.3±1.9 ka BP Hel-TL04108). Because of the evident climate amelioration and development of vegetation, the sand drift was restricted only to the territory of the location of the large and relatively high dune complexes. During the Atlantic time only large dunes were active within areas of high dune concentration (dune complexes) or good natural drainage conditions, i.e. in the landscapes where sufficient dry sand amount for wind transportation was available (sites Markalni 8.5±1.4 ka BP Hel-TL04109, Sarkankalni 8.5±1.8 ka BP Hel-TL04134, Veprishi 6.4±1.6 ka BP Hel-TL04110 and 7.1±1.2 ka BP Hel-TL04111). As of the end of the Atlantic any wind driven dune activity was over in the Seda plain. For this time there is no evidence of the later large-scale dune movement reactivation in territory under consideration.

According to analysis of the long axis of the arrangement of the parabolic inland dunes and locations of the dune complexes, the main wind directions were from W to E and from WSW to ENE during the phase of the dune stabilization. Inconsistency of the parabolic dunes forms and grouping of dunes into complexes suggest variable main wind direction and/or uneven dune base forming drift sand/glaciolacustrine material surface. Reasons of lack of the inland dunes in the southern (located at the foot of the Vidzeme upland) part of the Seda plain could be explained by poorer natural drainage conditions or high groundwater table.

As the definite age of Smiltene and Strenči glacial lakes is assumed only tentatively (Āboltiņš, 1971; Stelle et al., 1975), the oldest OSL ages of the inland dune sands can be used to set a minimal age of some phases of these basins. The dune base height at the Silezers sampling site is 54.5 m a.s.l and the corresponding OSL dating is 11.8±2.4 ka BP (Hel-TL04144) thus it can be concluded that any proglacial or remnant lake with the water level above 54 m a.s.l. is older than 9.4 ka BP. Also taking into account the Kauci_01 dune sand sample age (11.9±2.7 ka BP Hel-TL04142), the Smiltene ice-dammed lake and the uppermost levels of the Strenči proglacial lake can be assumed to be at least of the Younger Dryas age, but more likely taking into account other regional studies (Kalm, 2006; Zelčs, Markots, 2004) – considerably older, probably the Bölling or even Older Dryas time.

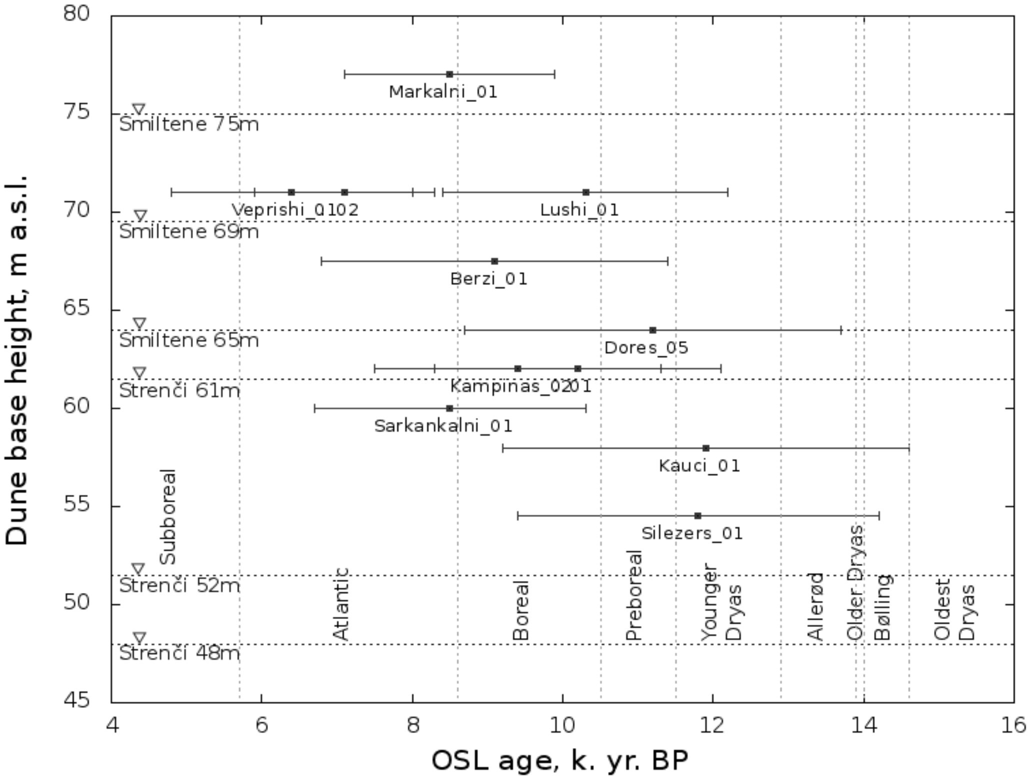

Fig. 1. Age of the dune sand samples and dune base height at sampling sites. Dashed lines and triangles denote main levels of local ice-dammed lakes.

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STOP 9:

SEDA MIRE – POST-GLACIAL PALUDIFICATION AND DEVELOPMENT OF MIRES IN LATVIA

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About 10% (6401.7 km^2) of Latvian territory is covered by wetlands. Wetland landscapes in Latvia are usually overgrown by trees and shrubs due to draining of the area and the surrounding lands. Locally the wetland areas are used for agricultural purposes. Consequently, only 4.9% of total wetland area is in natural condition and according to the thickness of organic deposits and the trophic state, the wetlands may be divided into fens, transitional mires and raised bogs.

Paludification and the mire development in Latvia started at Late Glacial – Early Holocene time. At the beginning of Allerød, ca 11,800 ^{14}C yr BP, the territory of Latvia was completely free of continental ice. The accumulation of gyttja and fen peat started only after amelioration of the climate at the fall of the Younger Dryas. The oldest peat deposits are reported from the overgrown small lake depressions, including small kettles of glaciokarst origin on the uplands. In lowland areas paludification processes started later in very wet depressions. The oldest peat deposits contain remains of plants that grew under the excess of humidity: *Hypnum* moss and different sedges (*Carex lasiocarpa, C. appropinquata* and *C. teretiuscula*), together with *Sheuchzeria* and horsetail (Kalnina 2007). Trees and shrubs were very scantily represented by dwarf-birch and birch. Following the decay of these plants, a variety of peat types were formed (*Hypnum*, sedge-grass, sedge-grass-*Hypnum* and less frequently wood-sedge-grass and wood).

The Seda Mire in northern Latvia is one of the largest mires in Latvia. According to pioneering studies by P. Nomals (1938), the mire covers an area more than 9000 ha. According to the data of Latvian Peat Fund (1980), peat deposit of Seda Mire (No. 2409) cover an area of 7252 ha, of which ca 6300 ha is covered by fen peat. The mire is located in the lowest part of the Seda Plain, which fills a shallow bedrock depression. Due to drumlinization and later aeolian activity, the modern terrain of the Seda Plain is slightly undulating. Fine-grained proglacial deposits of the Late Weichselian Strenči local ice-dammed lake form the mineral base of the Seda mire. Geological setting and topographic conditions favored the formation of the mire. The latter developed partly due to overgrowing of small lake basins, partly due to paludification of mineral soil. Initially the drainage of the Seda Plain was towards the Seda River. Later, due to peat formation in the central part of the present mire, a watershed between the Seda and the Gauja River basins was gradually formed. Fens and transitional type of mire dominate with comparatively flat topography at an altitude of 45–47 m a.s l. Raised bog occupies the central part of the Seda Mire and its convex surface raises few m (51 m a.s.l.) above the surrounding flat mire terrain. Small hollows, pools and lakes characterize the lanscape of the raised bog.

Palaeovegetation studies of Seda Mire deposits indicate that the peat accumulation took place since the Boreal Time (Fig. 1). Macroscopic remains of sedge, grasses and seeds were found in the peat. Pollen data and ^{14}C dating indicate fen and transition peat development from the Boreal until middle Subboreal, when a rapid changes in climate and peat formation conditions occurred. Formation of the fen type grass and grass-sedge peat occurred during the Holocene climatic optimum (Atlantic Time), when broadleaved forests spread in the surroundings of the mire (Fig. 1).

Table 1. General characteristics of different mire types in the Seda Mire.

Mire type	Area (ha)	Medium thickness of peat layer (m)	Largest thickness of peat layer (m)
Raised (moss) bog	1466	4.00	7.3
Transition mire	3824	3.01	6.0
Fen (grass) mire	1656	3.35	5.8

The most favorable conditions for the forest overgrowth almost over the entire territory of the Seda Mire prevailed during the middle part of the Subboreal time. This event is clearly marked by a layer of wood remains, which covers the woody peat layer dated to 4080 ^{14}C yrs BP. Climatic conditions were evidently sufficient for tree growth, as the largest pine stumps refer to approximately 60–80 years old trees. About 2800 yrs BP, a rapid climatic change took place, and in large areas wood layer became covered by a low decomposed peat of raised bog. The latter is represented by cotton grass-*Sphagnum* or *Sphagnum* peat. Pollen data show significant number of cultivated land and ruderal plant pollen in the diagram interval, corresponding to the Atlantic Time. This indicates the presence of a Stone Age man in the surroundings of the Seda Mire at that time.

At present Joint Stock Company "Seda" is cutting peat in almost all over the fen-type peat distribution area. The excavated area has been left for mire re-naturalisation. A large part of it is normally flooded and mixed swamp forest with reeds and shrubs will be restored. Mosaic landscape of large ponds, shrubbery and overgrown millet peat fields is expected to be formed. The restricted nature reserve area „Seda Mire" (7240 ha) was established in 1999, with the aim to protect important biotopes for water bird nesting.

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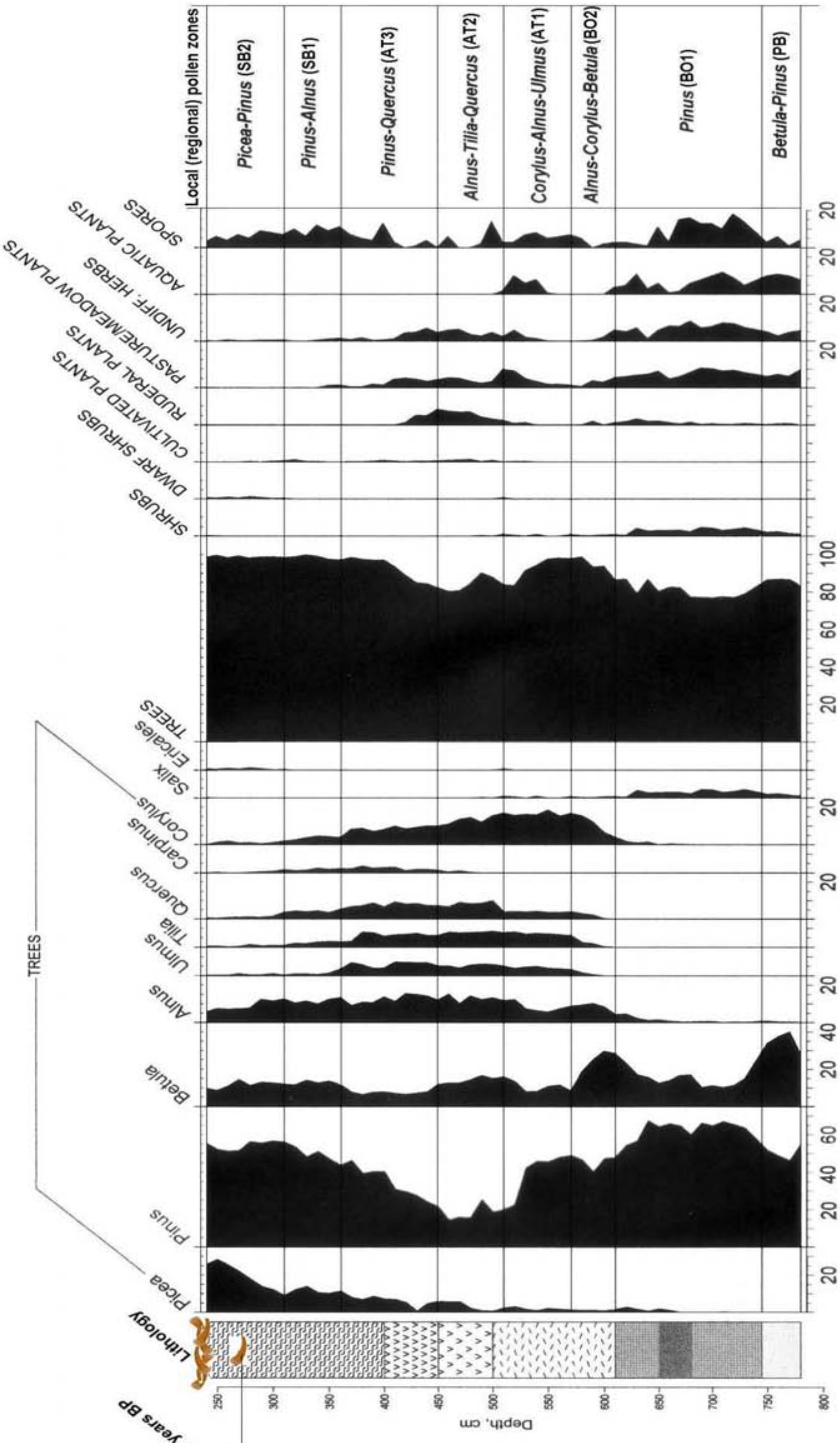

Fig. 1. Pollen diagram from the lower portion of the Seda Mire deposits (2.5 m of raised bog peat layer is not represented)

STOP 10:
KONUKALNS MEGADRUMLIN: INTERNAL STRUCTURE AND TIME TRANSGRESSIVE DEVELOPMENT OF DEFORMATION STRUCTURES

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The Ķoņukalns megadrumlin is the largest drumlin of Latvia. It is located on the NE side of the Burtnieks drumlin field between the Rūjiena town and the Latvian-Estonian border. This prominent ridge is about 10 km long, up to 2.5 km wide and 35 m high. It trends NNW-SSE with a slight curvature along its crest, convex to the SW side, and consists of several elongated segments separated by valley-like depressions. The largest valley-like depressions divide the megadrumlin into five main segments, but smaller ones cut into the slopes of the two largest segments. The depressions are 5–10 m deep, and they trend at an angle of up to 60º to the long-axis of the megadrumlin. Some of them slope upglacier, particularly in the NE part of the landform.

The overall stratigraphy of the megadrumlin was based mainly upon logs of test drillings and wells but detailed studies of the internal structure of the central highest segment were carried out in two large gravel pits.

General setting

The geological structure of the Ķoņukalns megadrumlin was already described by Dreimanis (1938), but the dominance of the lowermost greyish till in the core of the drumlin was exaggerated because the well drillers had interpreted the densely compacted and heavily deformed gravely sand as greyish sandy till. The latest test drillings, well logs and gravel pit sections data suggest the following pattern of the internal composition of the Ķoņukalns megadrumlin.

The lowermost structural and stratigraphic unit is the Middle Devonian sandstone bulging up under the southern part of the megadrumlin. A depression in the bedrock precedes the bulge at its upglacier end. Four till layers were encountered in the megadrumlin. The lowermost supposedly Middle Pleistocene till is grey and very densely compacted. The grey till fills in the small bedrock depression. The next three till units are reddish brown and belong to the Late Weichselian glaciation. Stratified drift layers, in places also by boulder concentrations and pavements, separate these tills. The lower reddish brown till bed occasionally is up to 15 m thick, and it forms an elongated palimpsest ridge in the northern and central part of the megadrumlin. This till is platy or fissile, contains intercalations of gravely sand and grey till. It is separated from the next, fissile reddish brown middle till unit by a gravely sand layer through most of the section, and a clayey-silt layer with disseminated plant remains in the downglacier part of the megadrumlin. These plant remains bearing sediments that also occur in cores or flanks of some other subglacial bedforms of the northern part of the drumlin field could be tentatively assigned to Middle Weichselian interstadial. The upper and middle reddish brown tills are separated by gravely and fine-grained sand, which forms an up to 20 m thick core of the main segment of the megadrumlin and is well exposed in all gravel pits.

Gravel pit sections

The gravel pits are located on the western slope of the megadrumlin. All sections are capped by 0.5–2 m thick upper till. In some places, this till layer may be subdivided into three subunits. The lower one has a distinct foliation structure, with some sand coating along the boundaries of plates. The middle subunit is massive or with thicker plates. The massive uppermost subunit contains many small lenses of sand or gravely sand. It could be interpreted as a subglacial melt-out till. In some parts of the sections, the lower or the lower and the middle subunits are absent. The till macrofabric measurements suggest the dominant stress direction from the inter-drumlin depression at the acute angle to the megadrumlin crest.

The glaciotectonically deformed gravel and sand core of the drumlin underlies the till. Its upper part is dominated by over-thrusted and sheared gravel and sand. The dips and their azimuths of the deformed gravel and sand beds were measured at some 30 places and the shear planes in sandy interbeds at a couple of places. Most of the thin gravel layers dip at 6º up to 15º towards NW, WNW and the cleavage planes at 65º–80º towards the NW, WNW.

Folding of sand and gravel had preceded the above discussed shearing. The dominant orientation of pebbles in recumbent folds, and the orientation of crests of folds (including minor drag folds) and several other sections suggest that the fold-forming stress came from WNW.

The sequence of sedimentation and drumlin formation

As it was mentioned above, the overlying clayey-sand and silt with plant remains were deposited probably during the Middle Weichselian interstadial. Gravely sand beds and lower reddish brown till were accumulated at the initial stages of the main Late-Weichselian ice sheet advance that was gradually replaced by vigorous erosion of the glacier bed. Both uppermost reddish brown till units were deposited and the accretional megadrumlin itself were formed later as a result of surges of the Burtnieks ice lobe.

In the Burtnieks field, drumlins are formed particularly on the upglacier slope of bedrock heights (Zelčs & Dreimanis 1997). A similar situation is also in the Ķoņukalns megadrumlin. Here the bedrock bulge of the Middle Devonian sandstone was the starting point (stable spot, according to Piotrowski et al. 2004) of the initial drumlinization process. The elongated shape of the bulge parallel with the ice flow direction, and the remnant of the grey till on its upglacier slope must have been produced by vigorous glacial erosion. The erosion was also followed by the accretion of the lowermost reddish brown till in a drumlin-like shape on the upglacier side of the bedrock bulge.

As the deposition of the sediments of the gravel, clayey-sand and silt with plant remains overlying lower reddish brown till most probably took place in the interdrumlin depressions, their present occurrence on top and downglacier end of the palimpsest accretional drumlin suggest their possible subglacial translocation during the next drumlinization process by the lateral thrusting from the interdrumlin areas. According to Zelčs and Dreimanis model (1997), at the beginning of aforementioned glacial surge the deposition of the middle reddish brown till produced the palimpsest drumlin.

The thick glaciotectonically deformed gravely sand, with various inclusions, which occurs between the middle and the upper reddish brown till, was emplaced there during the final drumlinization process as overthrusted sheets. The alternative hypothesis of its deposition by subglacial sheet floods in a drumlin-shaped glacial cavity (Shaw, 1983), mentioned in Zelčs and Dreimanis (1997) is unlikely, because the cavity erosion by a subglacial stream would also erode the crest of the underlying palimpsest drumlin. The final moulding of the megadrumlin took place during the North Lithuanian (Linkuva, Haanja) phase of the last glaciation, when the Burtnieks lobe surged in the Burtnieks bedrock depression for the last time. First the lateral stresses from the faster flowing interdrumlin ice folded and overthrusted sediments from the interdrumlin areas towards the axis of the megadrumlin, mostly in its main segment. Some local glacier and subglacial meltwater erosion followed, forming the diagonally downglacier trending valley-like depressions. Finally, the uppermost till was deposited by the glacial flow that gradually slowed down returning parallel to the axis of the megadrumlin.

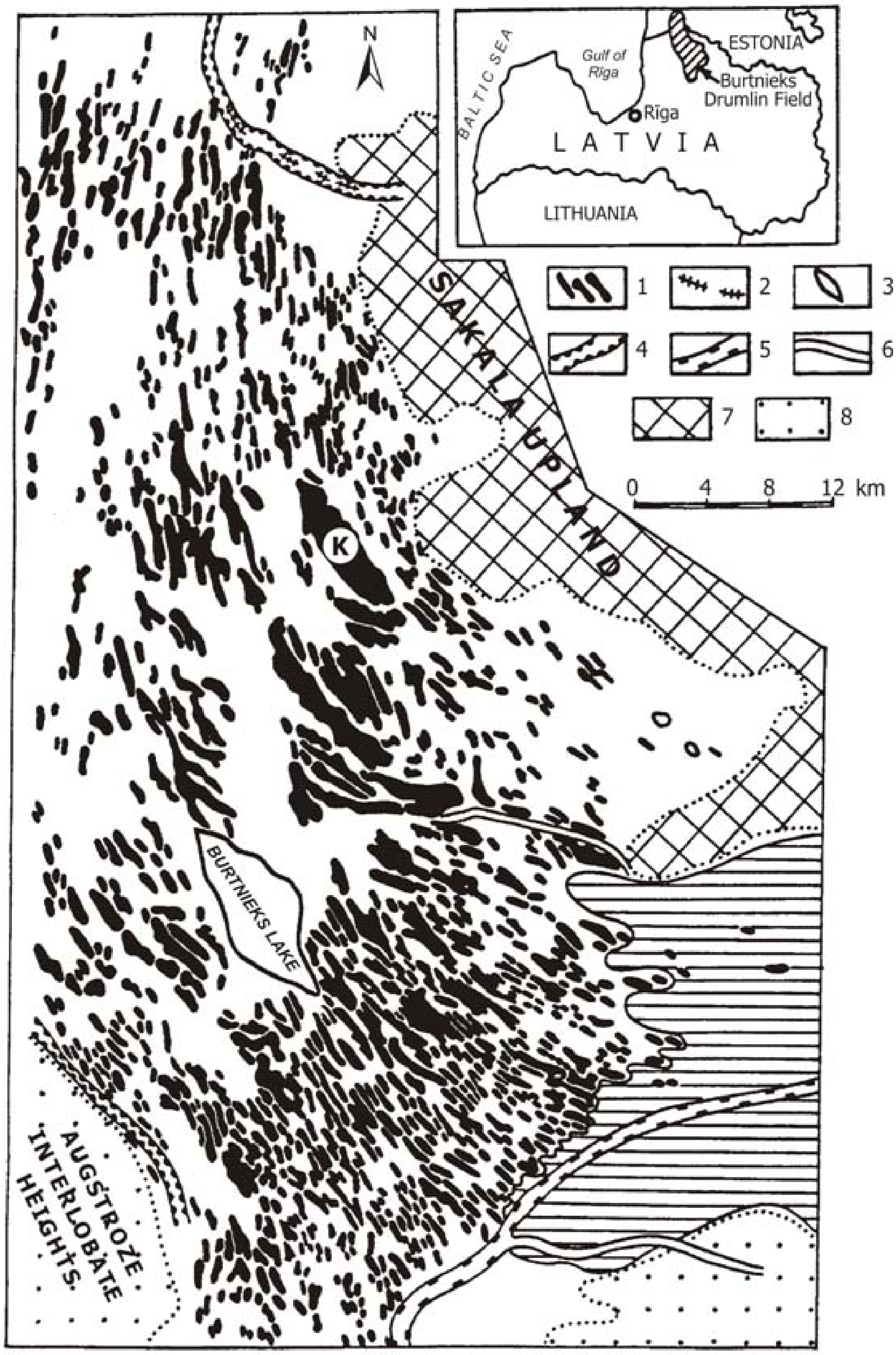

Fig. 1. Burtnieks drumlin field with respect to the sub-Quaternary surface features of surrounding regions (after Zelčs & Dreimanis 1997). Legend: 1 – drumlins; 2 – eskers; 3 – lakes; 4 – tunnel valleys; 5 – terraced drainage valleys of ice-dammed lakes; 6 – drainage channels of meltwater streams; 7 – Sakala bedrock upland; 8 – local bedrock highs; K – location of the Koņukalns megadrumlin and STOP 10. Horizontally hatched area marks the plain of the Strenči ice-dammed lake in northern Latvia

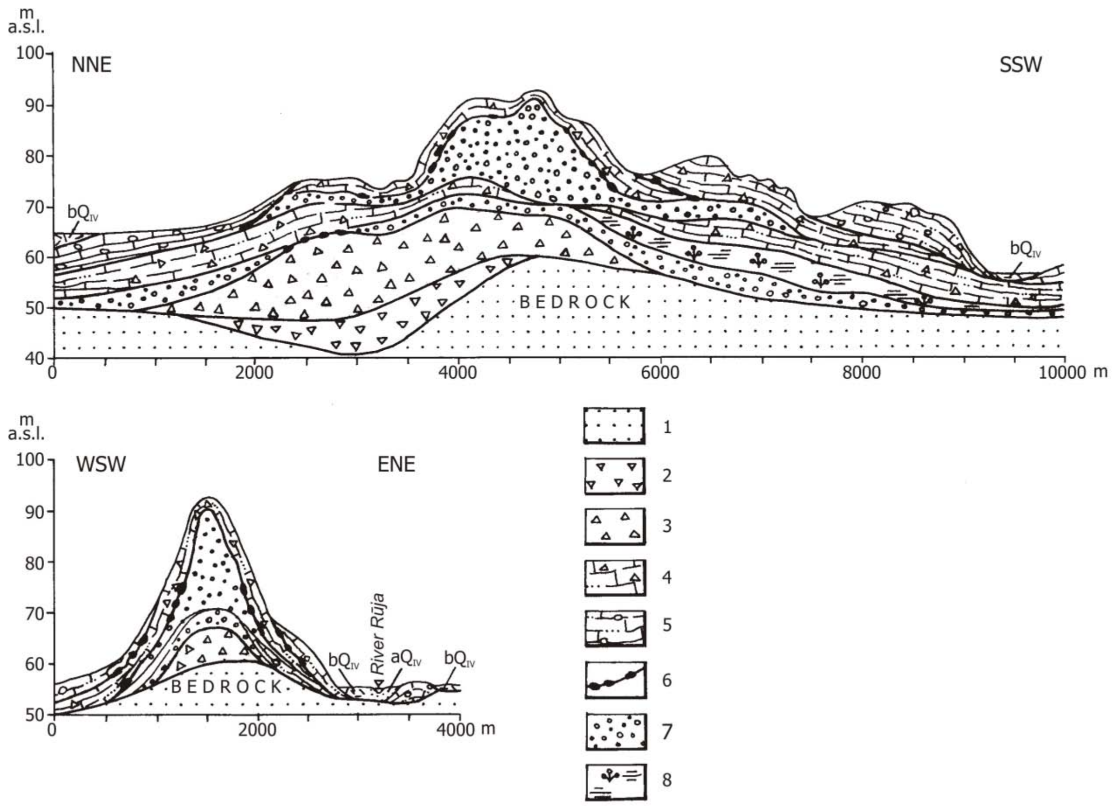

Fig. 2. Internal structure of the Koņukalns megadrumlin. Legend: NNE-SSW – longitudinal section, WSW-ENE – transverse section. 1 – Middle Devonian sandstone; 2 – lower densely compacted grey till; 3 – foliated reddish brown till with inculcations of grey till and gravely sand; 4 – upper fissile reddish brown till with thin lamina of gravely sand; 5 – upper weakly laminated reddish brown till with lens-like inclusions of gravel and sand; 6 – boulder pavement with striation; 7 – gravely sand with sand and pebble interlayers; 8 – clayey sand and silt with disseminated plant remains; bQ$_{IV}$ – Holocene peat; aQ$_{IV}$ – Holocene alluvial sediments.

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STOP 11:

HOLSTEINIAN INTERGLACIAL DEPOSITS AT KARUKÜLA IN SOUTHWESTERN ESTONIA

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Site of continental interglacial deposits, known under name „Karuküla“ is actually located in the Keskküla village at Kosenkranius (for now re-named to Aava) farmstead (Fig. 1). The site was discovered and thoroughly studied in 1940-s by prof. Karl Orviku. Karl Kosenkranius, the owner of the farmstead at that time, informed prof. K.Orviku about the peat and gyttja at the site, after he had read in newspapers the information on interglacial deposits at Rõngu. K.Orviku (1940) presumed a lacustrine origin of the silt and gyttja at Karuküla as fish scales and bones were found in the sediments. From 1940-s until 1980-s all together ca 70 boreholes and excavations were made by various investigators (K.Orviku, K.Kajak, E.Liivrand, A.Raukas, J.-M.Punning) at the site. The interglacial deposits are distributed in an area of ca 1 ha (Fig. 2) and are glaciotectonically deformed and probably of allochtonous bedding. Under the Karuküla site and to west of it is located a NE–SW-oriented buried Abja – Treimani valley that has cut ca 100 m into Devonian and Silurian bedrock. According to K.Kajak et al. (1970), there are at least four till beds in this buried valley and the oldest is presumably of Lower Pleistocene till. The Karuküla interglacial deposits, that consist of peat, compacted gyttja and silt with organics admixture, are covered only with thin (1–2 m) layer of Late Weichselian till (Fig. 3). Six main

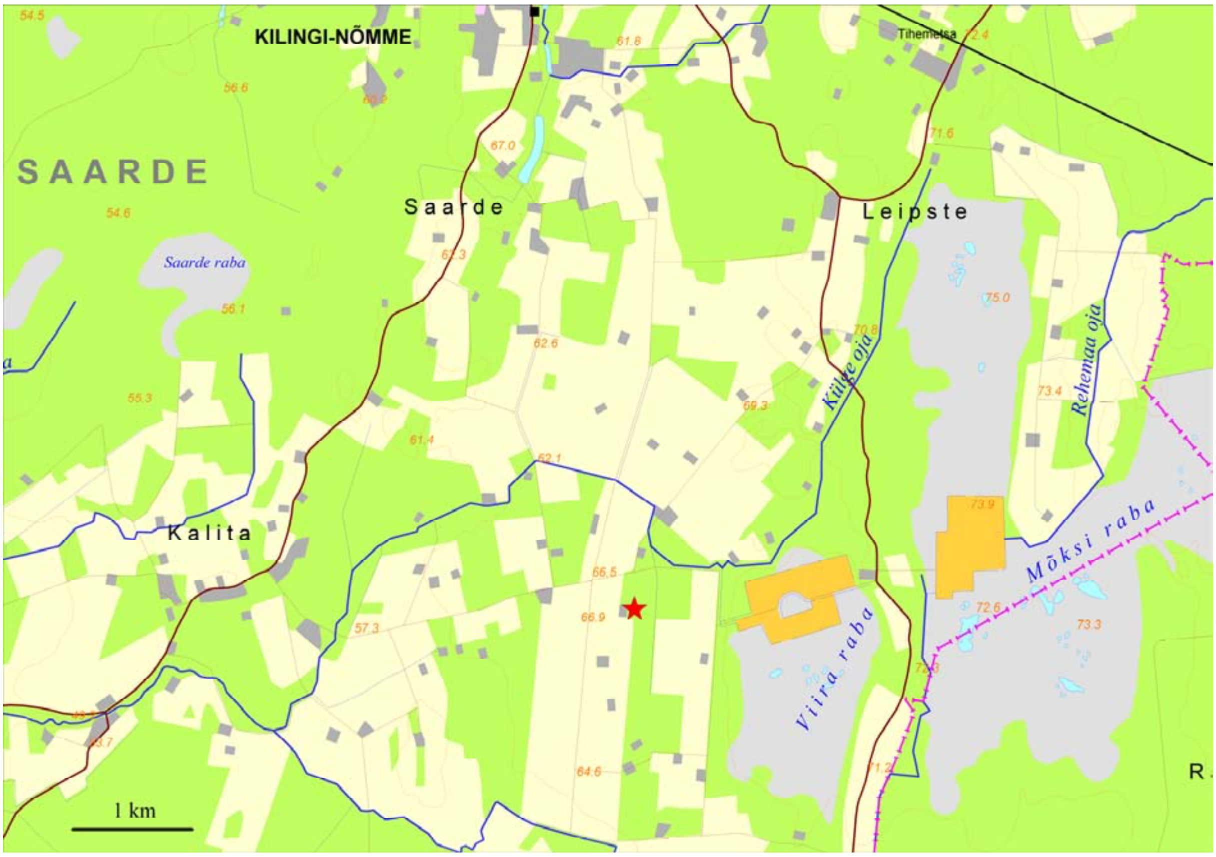

Fig. 1. Location of the Karuküla site of Holsteinian interglacial deposits.

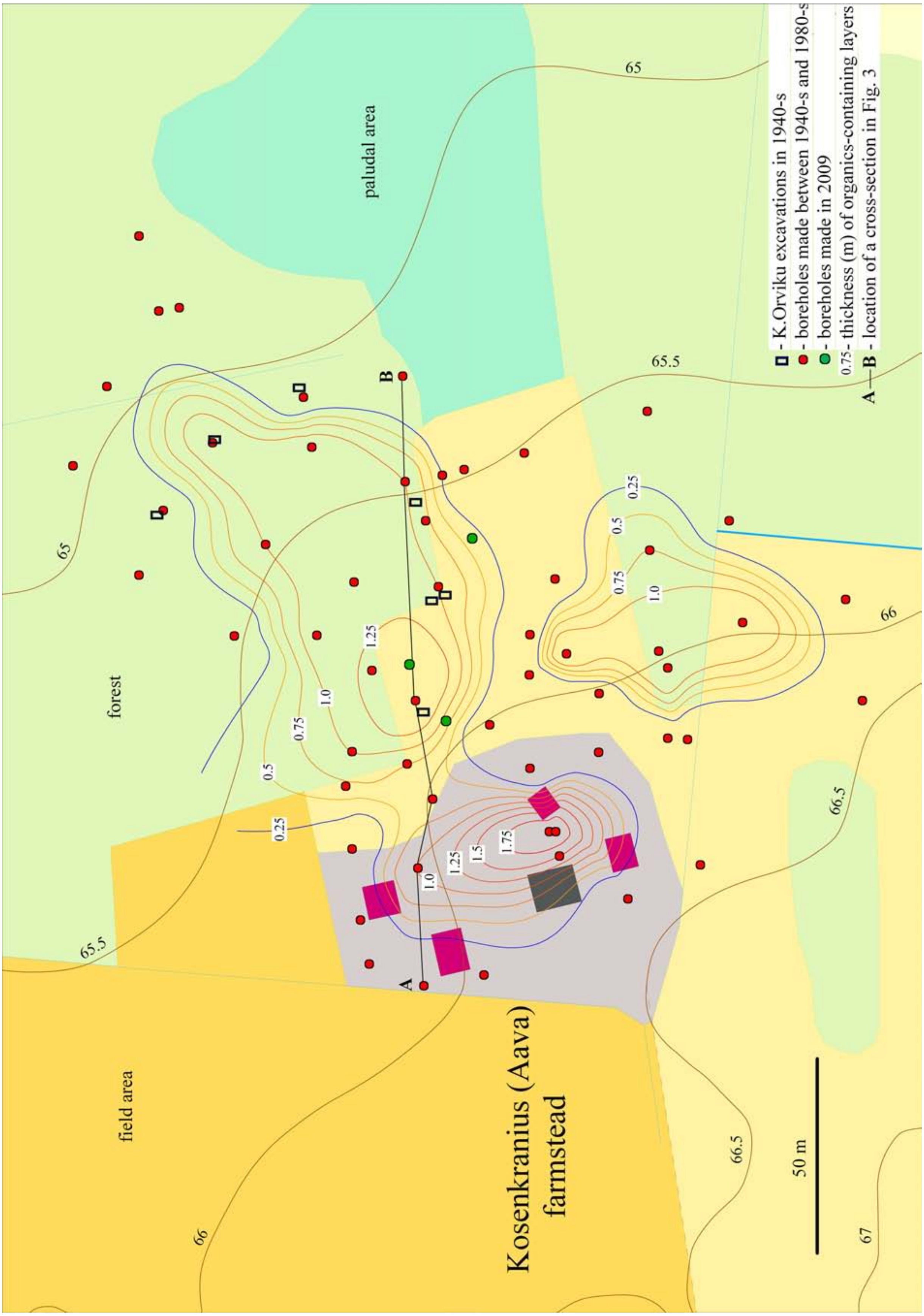

Fig. 2. Location of excavations and boreholes, and thickness of Holsteinian strata at the Karuküla site.

sedimentary units were identified in cross-section at the Karuküla site (listed from the ground, with thickness of layers in brackets; see also Fig. 3):

1. Reddish brown till with gravel and many erratics on the surface (1–2 m);
1. Stratified sand of varying grain size with inclusions of guttja, peat and wood fragments (0–1.25 m);
2. Brownish-black forest peat rich in pieces of wood (0.15–0.7 m);
3. Brownish-black to dark grey compacted gyttja, often brecciated into small debris (0.25–1.0 m);
4. Yellowish-grey silt and fine sand with thin organic-rich layers (0–0.4 m);
5. Yellowish sand and gravel, occasionally contains small pieces of gyttja (2+ m).

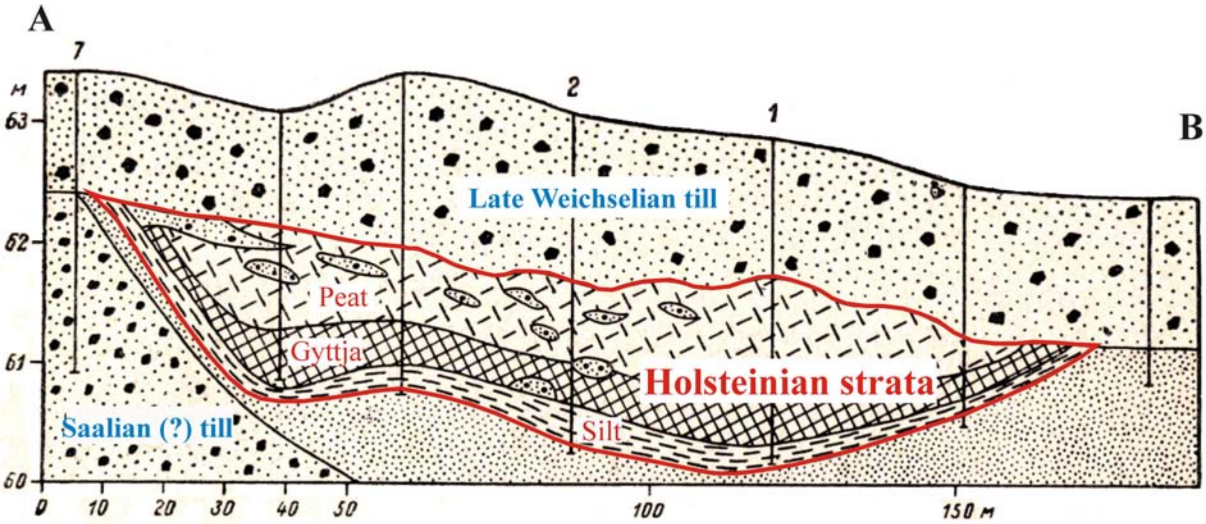

Fig. 3. Cross section at the Karuküla site, modified from Orviku & Pirrus 1965. See location in Fig. 2.

According to E.Liivrand (1991) there seem to be three large and two small lumps of interglacial deposits close to the surface under the till. Since the first studies in 1940 the Karuküla section had been correlated with the Riss-Würm Interglacial (Orviku 1944), the Brørup interstadial (Orviku & Pirrus 1965) and with the Middle Weichselian interstadial (Punning et al 1969, Serebrjannyi et al 1969). This last correlation had been supported by 20 ^{14}C dates from peat, gyttja and wood (ages between 33.45±0.8 and ≥58.6 ^{14}C ka BP; see Kalm 2005 and references therein).

The Holsteinian age of the deposits has been established through detailed palynological and carpological (plant macrofossil) investigations (Velichkevich & Liivrand 1976, Liivrand 1984, 1991). Palynological analyses done by E.Liivrand resulted in a diagram with the following pollen zones (Fig. 4):

I – lower part of compacted gyttja – *Betula* and *Pinus*;
II – upper part of compacted gyttja – *Picea* and *Alnus* maxima;
III – lower part of (*Phragmites*) peat – *Tilia*, *Quercus* and *Ulmus* maxima;
IV – lower part of forest peat – *Picea*, *Abies* and *Carpinus* maxima;
V – upper part of forest peat – *Betula* and *Pinus* with *Picea*.

According to E.Liivrand (1991), this pollen diagram is complete and the determined pollen zones correspond to the pollen zones of Holsteinian deposits in eastern Baltic region.

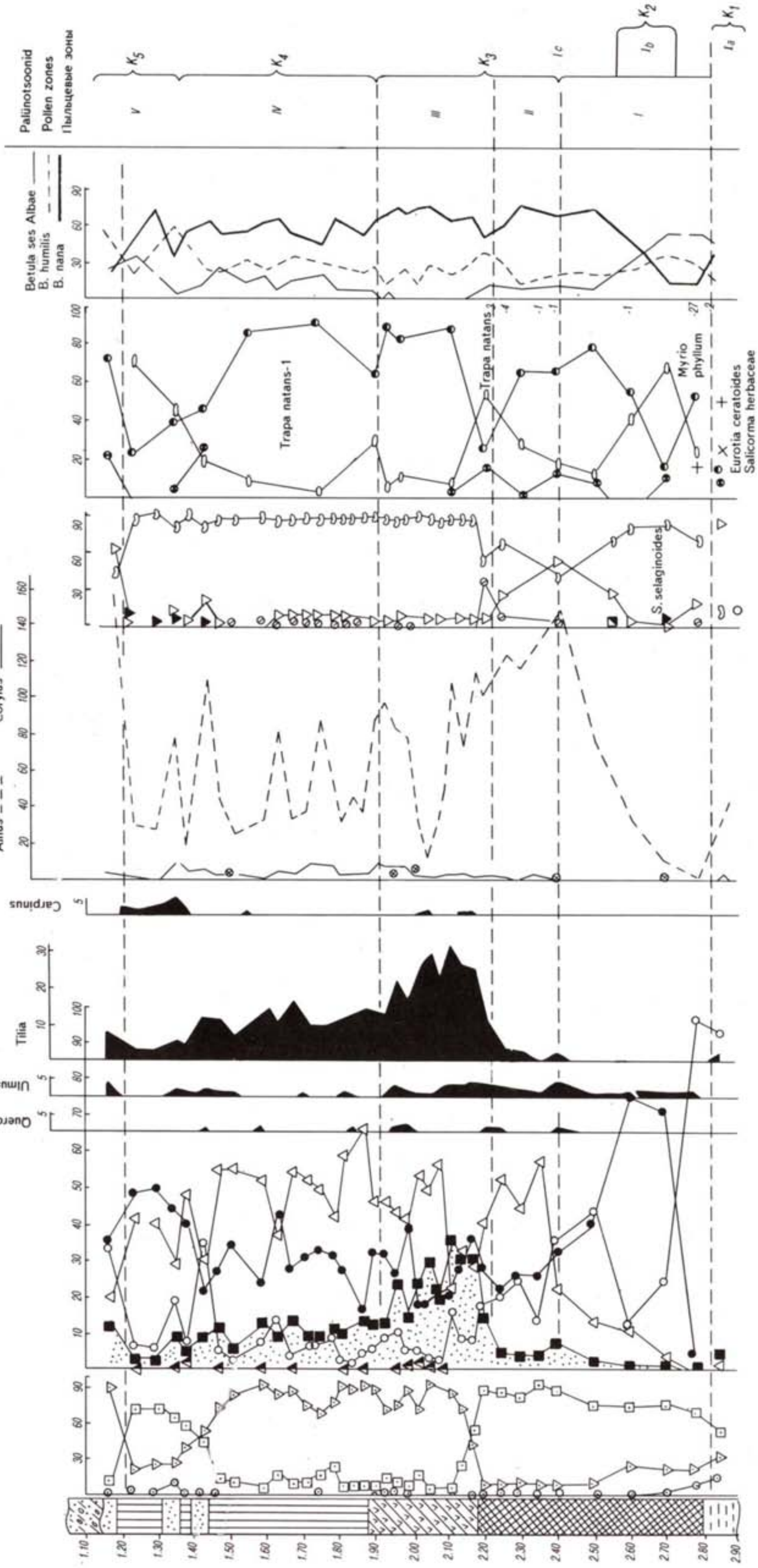

Fig. 4. Palynological diagram of the Holsteinian interglacial sediments of the Karuküla section (after Kajak 1995, analysed by E.Liivrand)

In 2009, we drilled at the site few new boreholes and recovered the peat and gyttja layer in thickness between 0.38–1.14 m. New core samples of peat and gyttja for U-series dating, and silt for OSL dating were taken in order to affirm or disprove the pollen-derived Holsteinian age of the sediments.

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STOP 12:
LATE SAALIAN CLAY SECTION AT ARUMETSA, SOUTHWESTERN ESTONIA

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The Arumetsa site (58°04´00´´N, 24°31´40´´E) is located in southwestern Estonia, about 40 km south of Pärnu town and 3 km from the coast of the Gulf of Riga (Fig. 1). The clay at Arumetsa is an important raw material for production of lightweight expanded clay aggregates (ECA) in the near-by factory, which is operated by the *Maxit Group AB* Company. The curiosity is that in

Fig. 1. Location of the Arumetsa site (A) and the clay quarry in 2001 (B) and in 2005 (C).

the earlier studies, related to the prospecting of clay resources, the clayey sediments at Arumetsa were supposed to be of Middle Devonian age. The clay complex was interpreted as a syngenetic clay lens within the reddish-brown Devonian silt- and sandstone, or as a fill-in of a river valley of that time. However, recent studies of clay mineralogy, microfossil and pollen assemblages, deformations in soft sediments (Rattas et al. 2001, 2004), and most recently ^{14}C and OSL age-determinations confirm the Pleistocene age of the sediments under discussion.

Regional geology at the site

The bedrock surface in the area is gently rising from –50 m a.s.l. at Kihnu Island in the Gulf of Riga, to about current sea level (0 m) near the coastline at Häädemeeste and up to 50 m a.s.l. on the slope of Sakala Upland in the east (near the Kilingi-Nõmme town). The upper sequence of the bedrock is composed of the Devonian sedimentary rocks (sand- and siltstones, dolomitic marls and clays) in thickness of ca 108 m at Häädemeeste and Arumetsa. Only the bottom of the deepest buried valley has incised into the Lower Silurian marlstone at a depth of –142 m a.s.l. at Abja-Paluoja and at –122 m at Treimani (Fig. 1).

The Quaternary cover is rather thin in the region and normally varies from <1 m to 50 m. Most of it is composed of Late Weichselian till and glaciolacustrine sand, silt and varved clays of the Baltic Ice Lake, and the Holocene marine and aeolian deposits. The greatest thickness of the Quaternary deposits occurs in buried valleys, reaching ca 200 m in Abja-Treimani valley (boreholes No 177 and 178 on Fig. 1). The boreholes revealed a sequence of Saalian (Ugandi) and Elsterian (Sangaste) tills and related aquatic deposits, but no interglacial sediments (Väärsi & Kajak 1969, Raukas 1978). On Kihnu Island and in its surroundings the up to 60 m thick Pleistocene sequence contains inter-till layer of dark brown to grey clay, that may reach thickness of 20 m. E.Liivrand (1991, 2007) identified Late Saalian and Early Weichselian pollen zones in this clay complex in the borehole No 526 on Kihnu Island. However, in the neighbouring core-section (No 21) almost complete succession of pollen zones from Late Saalian through the Eemian to the Early Weichselian was determined by L.Kalnina (1997, 2001). In the borehole No 25 in the Gulf of Riga the Early Weichselian sediments lay directly on the Devonian bedrock and no Eemian or older deposits were found (Kalnina 1997).

Morphology of the Arumetsa bedrock valley

The valley that hosts the Arumetsa clay deposit is an N-S oriented ca 2 km long and 300 m wide arc-shaped channel (Fig. 2). 80 m deep valley in the bedrock is incised into the reddish-brown weakly cemented sand- and siltstone and dolomitic marl of the Middle Devonian Aruküla and Narva Formations (Fig. 3). Cross-section of the valley is usually U-shaped with slopes of 20–30°, but in some segments even ≥50°. Western slope of the valley is higher (14 m a.s.l.) and steeper than that of the eastern slope (0–10 m a.s.l.). The longitudinal profile of the valley shows an undulating bottom with hollows, thresholds, and varying gradients, which is presumed to be the result of an irregular erosional cut. The irregularities in the valley bottom gradients indicate that the water flow may have been under substantial glacio-static pressure under the overlying ice sheet, confirming the theory of subglacial formation (ÓCofaigh 1996) of the valley.

Sedimentology of the valley infill

Sediments were described in detailed in the central part of the valley in depth down to 21 m. Information on sediments in the deeper parts of the sequence is obtained from drill core data. The entire open section of the clay and silt shows many soft-sediment deformation structures at different levels. They occur in a large variety of morphologies and older deformations are over-printed by younger ones. Slump-, rolling-, folded-, fault- and compaction structures are present in the laminated clay at Arumetsa. Three major sedimentary units are present in the section: 1) upper diamicton with fine-grained cover sand, 2) grey silt, silty clay and clay, and 3) brown to greyish brown silty clay and clay (Fig. 3). Each of the sedimentary units consists of few sediment facies.

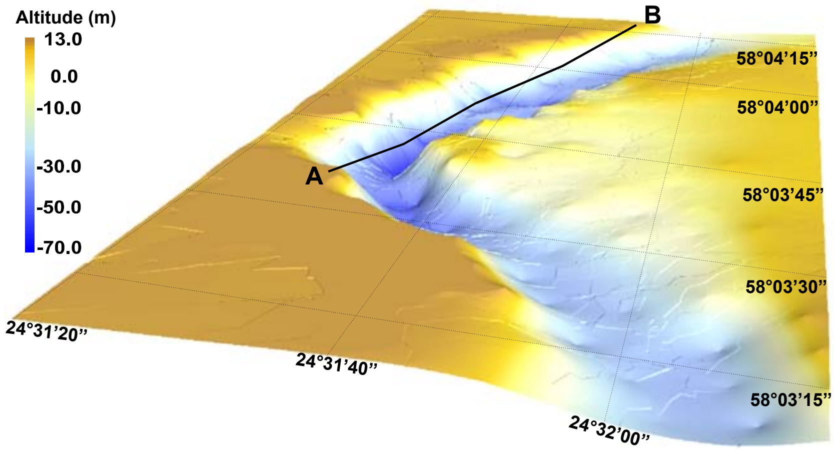

Fig. 2. DEM of the bedrock surface at the Arumetsa site.

Sedimentary unit 1

Cover sand

Fine sand and sandy silt occur as a thin (±1 m) continuous sub-horizontal sheet on top of the older sediments or bedrock. It is interpreted as the Late Weichselian lacustrine sand deposited in a shallow zone of the Baltic Ice Lake.

Diamicton

This facies includes a brownish, discontinuous, massive, medium-grained sandy-silty, matrix-supported (poor of clasts) diamicton in thickness of 2 m as a maximum. The diamicton has a sharp conformable lower contact with the silty clay (Unit 2) below. Beyond the bedrock valley, the diamicton lies directly on Devonian sandstone. Contact between the upper sand layer and the diamicton is transitional and in core sections often unclear. The diamicton is interpreted as the lodgement till of the last (Late Weichselian) glaciation. The waters of local- and the Baltic Ice Lake waters have later reworked its uppermost layer.

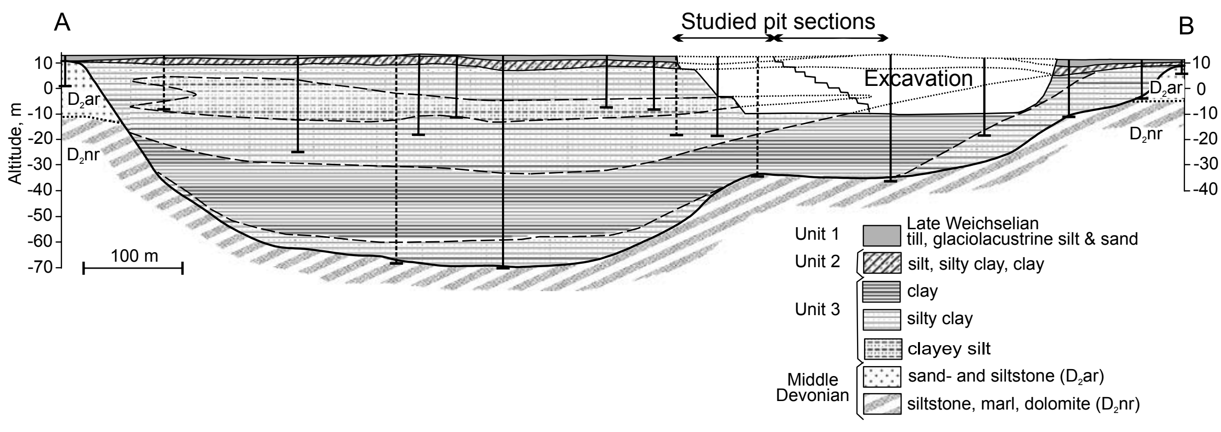

Fig. 3. Composite cross-section through the Arumetsa bedrock valley. Location of cross section see in Fig. 2.

Sedimentary unit 2

This unit of grey fine-grained sediments (fine sand to clay) represents a continuous depositional sequence and consists of two low-energy lacustrine facies. The laminated fine sand and silt facies conformably overlies the grey homogeneous clay, representing coarsening-upward succession. Thickness of the Unit 2 varies between 1 and 8 m, being greater in the central part of the valley. The whole sedimentary unit is glaciotectonically deformed and the orientations of large folds and fractures suggest the ice pressure from the north.

Laminated fine sand and silt
Grey, laminated and contorted silt is covered by up to 3 m thick yellowish grey fine sand and sandy silt, which has a sub-mm-scale organic-rich interlayers. The fine sand and sandy silt are well sorted but more-or-less regular lamination is heavily disturbed by sedimentary and glaciotectonic deformations. Variety of structures, including graded bedding, convolutions and load structures, and fold and drug structures are present.

Massive silty clay and clay
The vaguely laminated silty clay at the base of the unit gradually changes into homogeneous massive clay with peculiar crumbly or brecciate texture. Thickness of the clay facies reaches 3 m. There is a 1–3 cm thick layer of fine gravel at the base of the Unit 2 and a sharp undulating basal contact with the Unit 3 below. Character of the contact between the sedimentary Units 2 and 3 indicates a possible hiatus at that level. Few pieces of wood were found and ^{14}C dated from the base of this sediment facies.

Sedimentary unit 3

This sedimentary unit consists of two laterally discontinuous facies – the laminated silty clay and the homogeneous massive clay – and fills in most of the bedrock valley depression. The unit represents a long period of relatively steady para-pelagic sedimentation. Few dropstones (≤6 cm in diameter) of crystalline rocks were found in these sediments. Current lateral discontinuity of the unit is probably due to post-depositional deformation and erosion processes. Deformation structures occur in a large variety of morphologies and represent complex structures from deformations of different events in different times.

Laminated silty clay
Clay and silt rhythmites dominate with occasional ripple cross-laminated fine sand layers in the sediments. Typical silt and clay varves with distinct grading are present in rhytmites. Thickness of individual laminae is normally between of a sub-millimetre to few centimetres. Sand laminae and ripple-marks indicate turbidity and/or current processes during the sedimentation. Lamination in this facies is also heavily disturbed. Different soft-sediment deformations occur at various levels and scale, either within a single layer or in thicker bed complex. Deformation mechanisms and driving forces are related essentially to gravitational instabilities, compaction, dewatering and liquidization. Small-scale slumps and slides are probably resulted by underwater gravity sliding on the bottom of water body (lake). However, there is no evidence of mega-scale slumping from the surrounding valley slopes. Nevertheless, several erosional contact surfaces between different clay blocks may be interpreted as possible slide surfaces.

Massive clay
The laminated silty clay lithofacies is inter-bedded with thick (up to 10 m) layer of massive clay. The massive clay contains few, laterally discontinuous, thin (0.1–1 cm) layers of clayey silt. Locally globular pebble-size clasts of coarser soft sediments (silt, clayey silt) occur in the clay.

Deepest horizons of the Arumetsa clay section have been studied only from the core sections. According to this information a massive brown clay and clayey silt fills in most of the bedrock valley below the level of opened outcrops. At the very bottom of the Arumetsa bedrock valley, there is a 2 m thick, well sorted, unconsolidated layer of bright brown to brownish grey silty sand, which has sharp (erosional?) contact with overlying brown clay.

Microfossil and pollen record from Arumetsa clay

Poor in terms of quantity and not well-preserved ostracod, diatom and pollen material is characteristic to the Arumetsa clay.

Ostracod carapaces and their fragments are found from the upper part of the massive brown clay of Unit 3. Identification of ostracod species was complicated because the majority of the carapaces were broken. Only two species were identified and *Cytherissa lacustris*-dominated fauna characterize the ostracod assemblage with very few *Ilyocypris bradyi* individuals. Presence of *C. lacustris* refers to cold and well-oxygenated water in a freshwater lake. However, the species is tolerant to salinities up to 1.5‰ and to great thermal fluctuation (Meisch 2000), there is a possibility that also slightly brackish-water conditions may have occurred. The Late Saalian brackish-water phase has been described in marine sediments of Prangli section in northern Estonia (Liivrand 1991) and in the coastal areas of Latvia (Kalnina 2001). *I. bradyi* inhabits muddy and sandy bottom sediments of springs, ponds and lakes fed by springs. Thus, it indicates also cold-water aquatic environment, but of low water level. Usually *I. bradyi* is associated with very low salinities, but tolerates oligohaline waters up to 4.4‰ (Meisch 2000). The occurrence of *I. bradyi* indicates a gradual change to shallower water, which may be due to a cessation in the global sea-level rise and a continued isostatic uplift following the Saalian glaciation. The valley basin was probably isolated from the Baltic Sea basin and changed to a continental oligohaline lacustrine environment, which at least in first stadium was rather cold.

Diatoms were found only in the grey silt and silty clay of the Unit 2. Within the unit, their number and diversity are increasing upwards. The diatom flora is a typical fresh-water (oligohalobes) assemblage, dominated by planktonic species of genuses *Aulacoseira, Stephanodiscus, Cyclosyephanos* and *Cyclotella*. Spieces *A. islandica*, *A. granulata* and *St. astraea* represent about 70% of diatom flora. The occurrence of the marine centric diatom *Paralia sulcata* indicates probably to some open-marine influence that might be drifted to the site by currents or storms. Brown clay and silty clay of the Unit 3 is barren of diatoms. Probably the environment during the formation of the Unit 3 was not suitable for the diatoms. They need for their normal life more sunlight and clear water than may have been available in this suspension-rich body of water at that time.

Pollen of trees, mainly of conifers *Pinus* and *Picea*, is prevailing in the silty clay and clay of the sedimentary Unit 2. Among deciduous trees, *Alnus* and *Betula* dominate and only few pollen grains of broad-leaved trees – *Quercus, Ulmus, Tilia, and Carpinus* – were noticed. *Polypodiaceae* mainly represents herbs. The pollen, being mainly derived from trees, refers to their interglacial origin, but no any typical forest successions or pollen-zonal record of known interglacials were observed. Similarly to the Kihnu section (Liivrand 2007) the pollen record from Arumetsa may indicate redeposition from the Eemian sediments during the Early Weichselian time.

The Unit 3 is characterized by a very low concentration of not well-preserved pollen grains, which probably suggest redeposition. *Pinus* and *Betula* pollen were dominant, other trees (*Alnus, Picea, Tilia* and *Quercus*) and herbs (*Artemisia, Chenopodiaceae*, *Eurotia ceratoides,* and *Hippophae rhamnoides*) were represented only by few grains. Pre-Quaternary palynomorphs, such as *Gleichenia, Mohria, Eupicea, Trilobozonotriletes, Trematozonotriletes,* *Euryzonotriletes, Hymenozonotriletes, Veryachium* and *Acritarch* are relatively abundant in the Unit 3.

Age of the Arumetsa clay

Altogether six ^{14}C and ten OSL age-estimations (NLLD, Risø) have been performed from the materials available in the outcrop section. OSL datings of the laminated silty clay (Unit 3) yielded the ages between 120 and 151 ka that is the Late Saalian time.

OSL datings of the grey laminated fine sand and silt (Unit 2) gave the values between 37 and 40 ka, referring to the Middle Weichselian time of the deposition.

Five ^{14}C datings from pieces of wood, discovered at the boundary between the Units 2 and 3, support the OSL age of the Unit 2. Conventional ^{14}C analysis of the wood resulted in infinite values (>50 and >35.5 ka), but additional ^{14}C AMS datings (Hela) yielded finite ages between 40 and 44 ka BP. Another ^{14}C dating (4.2 ka) of organics from the cover sand confirms its Holecene age.

Two exceptional OSL ages (ca 230 ka) from the grey sandy silt layer (Unit 2) may equally indicate one of the three options – redeposition of the older material, incomplete bleaching, and poor luminescence properties – or their combinations.

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STOP 13:

COASTAL LANDFORMS AT RANNAMETSA AND TOLKUSE MIRE (SW ESTONIA): RECONSTRUCTING THE WATER LEVELS AND COASTLINES OF THE PAST BALTIC SEA

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Coastal landforms at Rannametsa and neighbouring Tolkuse mire are located in a flat coastal area of SW Estonia characterized by even topography and slow postglacial land uplift (about 1 mm/yr, Vallner et al. 1988). Even minor fluctuations in the Baltic Sea water level in this area have caused a notable shift of coastline and burial of the older deposits and prehistoric settlements. Geological data from Rannametsa and Tolkuse sites reflects the water level change history of the past Baltic Sea and this information will be discussed together with other shore displacement evidences from SW Estonian coast (Fig 1).

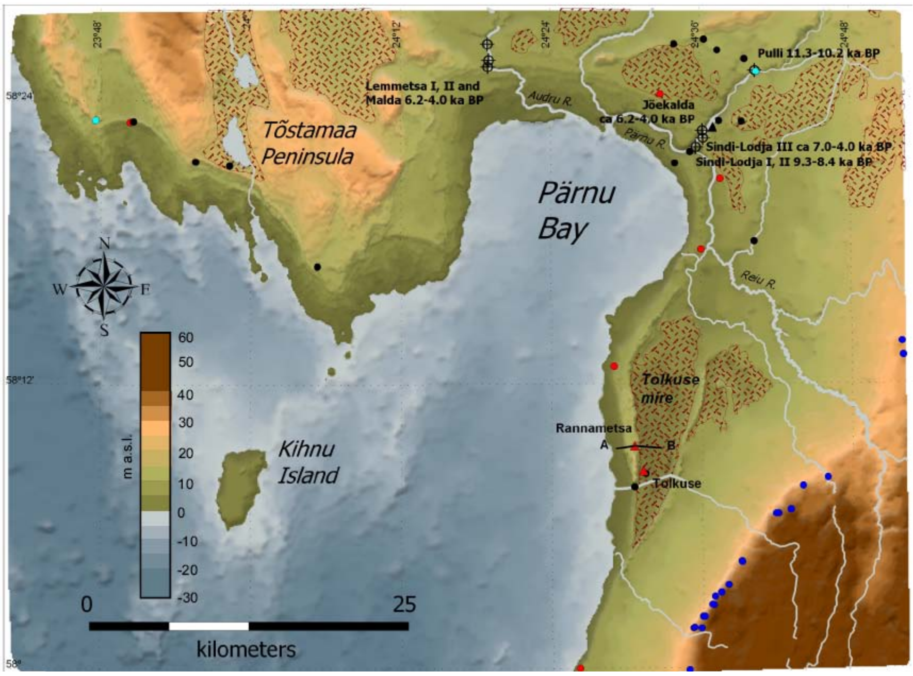

Fig. 1. Digital terrain model of the southwest Estonia showing the location of Rannametsa and Tolkuse sites (red triangle), and other investigated geological and archaeological sites, used for shore displacement reconstructions presented in Fig. 2. Sites with a buried organic matter and dated peat sequences are marked by black dots. Location of the Baltic Ice Lake (blue dots), Ancylus Lake (light blue dots), and Litorina Sea (red dots) coastal landforms are shown on the map. Brown hatching marks peat bogs.

Water level changes in the past Baltic Sea in southwestern Estonia

Reconstruction of a shore displacement of the past Baltic Sea is based on geological, geodetic and archaeological evidences. A new shore displacement curve was reconstructed recently for the SW Estonia on the basis of coastal landform elevations and ^{14}C dated buried peat and soil sequences below the transgressive Ancylus and Litorina sediments (Rosentau et al., subm.). The Holocene shore displacement curve for the area displays three regressive phases of the past

Baltic Sea, interrupted by the Ancylus Lake (10.6–10.2 cal ^{14}C ka BP) and the Litorina Sea transgressions (8.6–7.3 cal ^{14}C ka BP) with magnitudes of 12 m and 10 m respectively (Fig. 2). Coastal landform data in the area contains sites from the Baltic Ice Lake, the Ancylus Lake and the Litorina Sea stages (Fig. 1). Correlation of coastal landform elevations from the southwestern Estonia and neighbouring areas has been used to define the highest water levels and the shoreline tilting during those stages (Saarse et al. 2003). Shoreline data suggests the exponential-like decrease of shoreline tilting in time, reflecting slow-down of the land uplift. The tilting of the highest Baltic Ice Lake shoreline is 0.40 m/km^{-1}, being gradually lower in younger Ancylus Lake (0.27 m/km^{-1}) and Litorina Sea (0.13 m km^{-1}) shorelines. By combining the palaeoshoreline data with shore displacement curve and with relative sea level data for the last century (Vallner et al. 1988) a spatial-temporal water level change model was developed. This model was applied together with the digital terrain model to reconstruct the palaeocoastlines for the different Baltic Sea stages. This made possible a study of the relationships between coastal change and displacement of early prehistoric settlements in the area, including the oldest Estonian settlement at Pulli (dated to 11.2–10.2 cal ^{14}C ka BP).

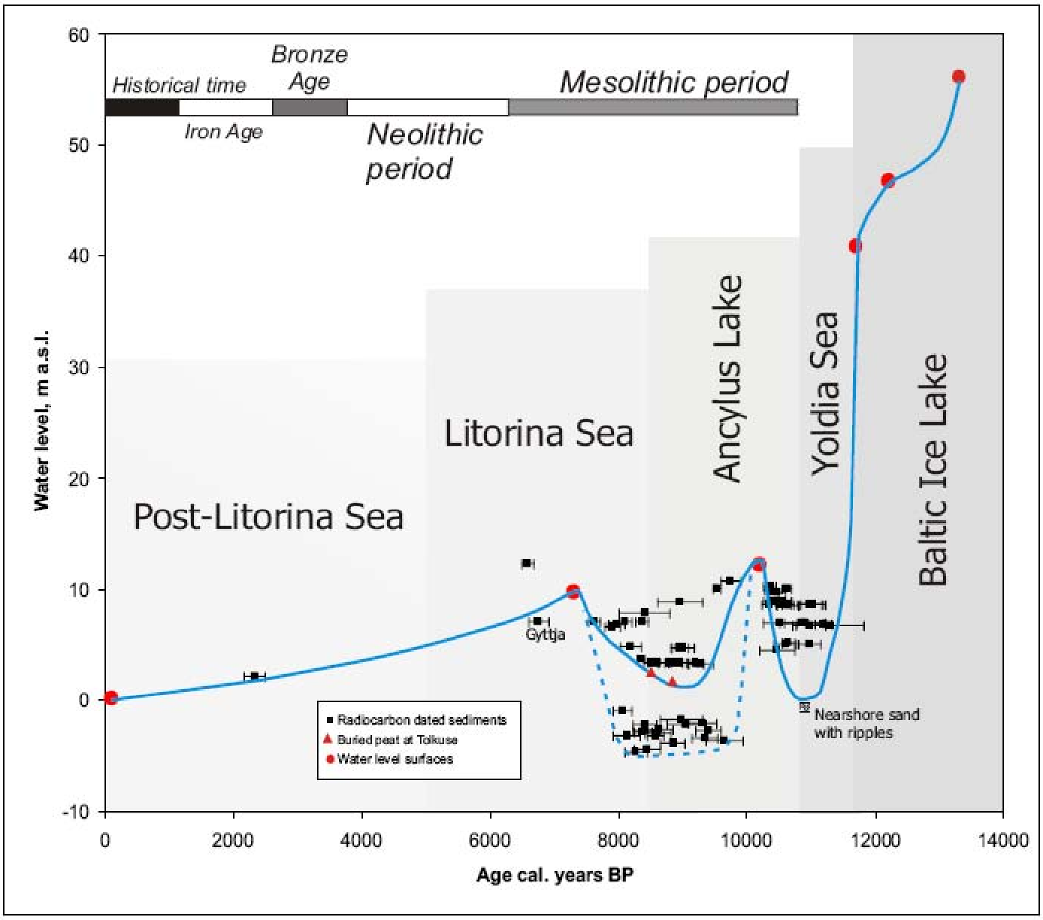

Fig 2. Shore displacement curve for the Pärnu area. All sites were corrected by spatial spread and referenced to the Paikuse location (black triangle in Fig. 1). Baltic Sea stages are according to Andren et. al. (2000).

Coastal landforms at Rannametsa and Tolkuse mire

Fieldworks in Tolkuse mire in summer 2009 led to the discovery of a peat layer buried by Litorina transgression (Fig. 3). The Tolkuse mire, located behind the Rannametsa coastal landforms of the Litorina age, was selected for investigation in order to examine the water level during the pre-Litorina transgression. ^{14}C dates from the buried peat in Tolkuse core suggest the water level low stand (around 1 m a.s.l.) at about 8.8–8.2 cal ^{14}C ka BP. The peat layer is covered with sandy deposits, which are alternating with thin peat layers and are characteristic transgressive sediments of the Litorina Sea, found elsewhere in southwestern Estonia. Sandy deposits are covered by gyttja which in turn is overlain by younger peat deposits. First ^{14}C dates from the upper part of the buried peat and basal part of the upper peat suggest that the lagoon/lake system existed in the area of present Tolkuse mire between 8.2–4.2 cal ^{14}C ka BP. Tolkuse lagoon was probably isolated from the Baltic Sea and a freshwater lake existed after the Litorina transgression between 7.3–4.2 cal ^{14}C ka BP. Further diatom analysis and ^{14}C dates from Tolkuse site are expected to determine an exact time of the isolation.

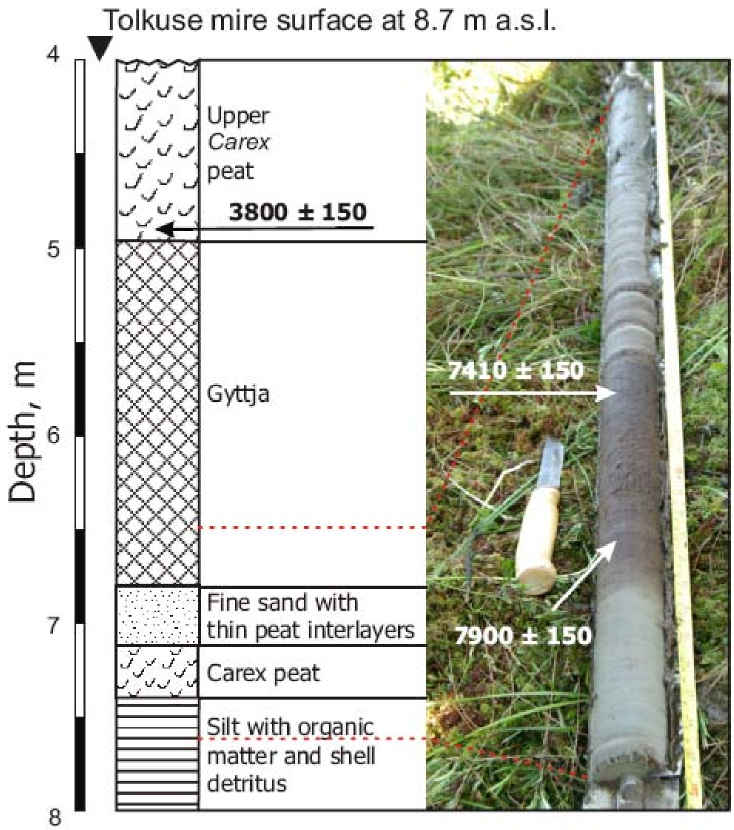

Fig 3. Lithology and uncalibrated ^{14}C dates from the Tolkuse core. Site location see in Fig. 1.

The development of the coastal landform system at Rannametsa is closely related to the development of the lagoon/lake system at Tolkuse. Coastal landforms at Rannametsa were first studied by H.Kessel (1965) and later by H.Hyvärinen et al. (1992). Hyvärinen et al. (1992) described a section of a peat and clayey gyttja at an altitude of 3–4 m a.s.l., which is buried under the sandy deposits of the Litorina Sea coastal ridge (Fig. 4). Earlier ^{14}C dates from Rannametsa and from Tolkuse sections presented here, suggest a simultaneous formation of the buried peat deposits in both mires at about 9.0–8.2 cal ^{14}C ka BP. Later, during the Litorina Sea transgression at about 8.2–7.3 cal ^{14}C ka BP the formation of a coastal ridge in Rannametsa took place and the lagoon was formed behind it at Tolkuse basin. The crest of the coastal ridge is at an altitude of 5.9 m and is interpreted to represent the highest Litorina Sea level in the area. Due to later aeolian activity, coastal dunes with relative height of about 29 m were formed on top of the coastal ridge described above. Coastal dunes at Rannametsa and Tolkuse mire belong to the Luitemaa ('Land of dunes') Nature Reserve.

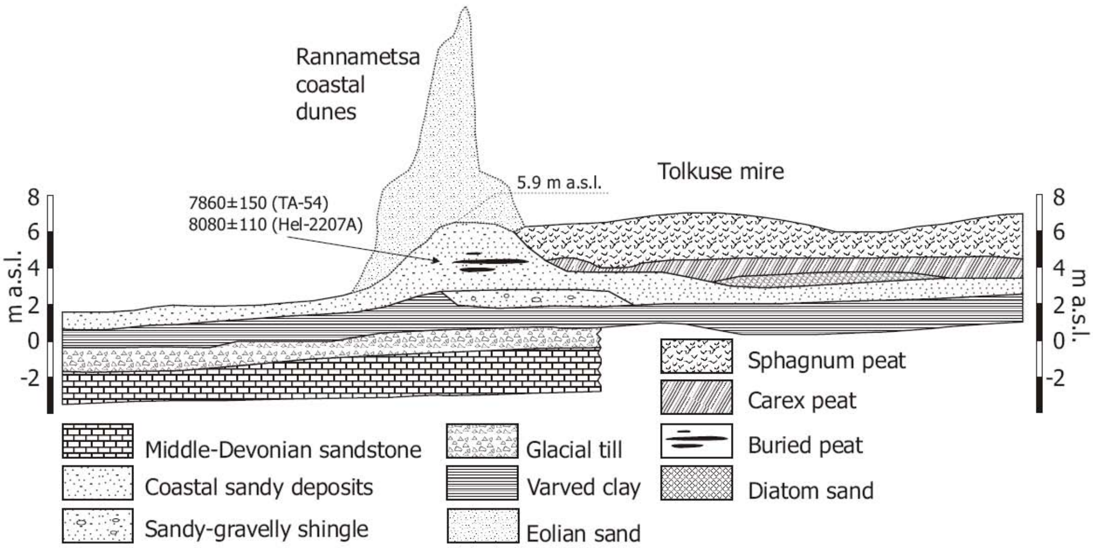

Fig 4. Geological section (A–B) through Rannametsa coastal landform system and the Tolkuse mire (modified from Hyvärinen et al. 1992). Location of the profile is shown on Fig. 1.

Coastal dunes are widely spread in the surroundings of the Pärnu Bay. Almost continuous chain of dune ridges starts from the Latvian coast in the south, close to Rannametsa splits into two – Ancylus and Litorina chains of dunes, and gradually disappeares near Uulu at the back of the Pärnu Bay. Interrupted parallel chains of dune ridges appear again at the western coast of the Pärnu Bay and follow the coastline of the Tõstamaa peninsula (Fig. 1). Highest dunes are located at Rannametsa but also at Tõstamaa the relative height of the dunes reaches 15 m. Described ridges follow the Ancylus Lake and Litorina Sea coastlines and similarly to Rannametsa site, they are complex forms where former gravely beach ridges are buried under the dunes.

Both, inland and coastal dunes are present in Estonia. Continental dunes occur in northeastern Estonia, north of Lake Peipsi. These are parabolic transverse formations, 0.8–2.7 km in length and up to 15–20 m high. Their north or northwest slopes are gentle (3–18^{o}) but leeward slopes steeper (18–24^{o}), indicating a westerly or northwesterly palaeo-wind direction during the dune formation. Continental dunes originate from Younger Dryas/Pre-Boreal period when a significant regression of the glacial Lake Peipsi took place in eastern Estonia. Coastal dunes occur along the ancient coastlines of Estonia and were mainly formed during the Baltic Ice Lake stage and during the transgressive phases of Ancylus Lake and Litorina Sea. Because of the land uplift, coastal dunes are located at some distance from the present coastline and at different altitudes. Depending on their closeness to the ancient shorelines, coastal dunes are traditionally classified as the Baltic Ice Lake, Ancylus Lake or Litorina Sea dunes. However, they may have been repeatedly reblown and therefore must be taken with caution in palaeocoastline reconstructions.

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STOP 14:
PROGLACIAL VARVED CLAYS AT PÄRNU: SEDIMENTARY EVIRONMENT, VARVE CHRONOLOGY, ENVIRONMENTAL HAZARDS RELATED WITH THE LANDSLIDES

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During the last Termination in front of receding ice margin an extensive proglacial bodies of water were present configuration and depth of which changed in accordance to the ice recession. After the ice retreated from western Estonia waters of the Baltic Ice Lake flooded the area. This is reflected in wide distribution of glaciolacustrine deposits including annually laminated varved clays (Fig. 1).

Distribution of varved clays in western Estonia, their main lithological characteristics and geotechnical properties have been studied since 60-ties of the last century. Latest studies have been concentrated to the part of proglacial basin located at the southern, distal slope of the Pandivere-Neva belt of marginal formations (ca 13500 yr BP). Over entire study area, the hummocky upper surface of the Late Weichselian bluish-grey loamy till is covered by glaciolacustrine varved clay or silt. Maximum thickness of varved clay is 30 m (in the city of Pärnu) being mainly 5–9 m. Locally rapid changes in clay thickness are due to underlying till topography. The upper surface of clay is dipping towards south being at an altitude of –8 m ca 5 km offshore and –2.5/–3 m at the current coastline of the Pärnu Bay. At the distance of ca 2 km north of coastline the altitude of clay surface is rapidly changing from zero to 4 m.

Clay fraction is dominant, particularly in the upper portion of varved clays where it constitutes 69–79%. The content of silt fraction increases and clay fraction decreases (down to 46–51%) towards the depth, which is explained with the increase in the thickness of silty summer layers. Within a single varve, the summer layer consists mainly of fine- and medium-grained silt (up to 90%). Clay fraction usually starts to dominate already in the upper portion of the summer layer and constitutes at least 70–80% of the winter layer.

Five sedimentary units of clay have been described according to the varve thickness, inside structure and fabric (Hang et al. 2007). Units A, B, C and E represent a varved complex displaying gradual thickening of varves downwards. In NW part of the Pärnu Bay and sporadically on onshore areas a massive silty clay unit (unit D) with dispersed sand grains and has been described within the varved clay complex. Its composition responds to clayey waterline glacial diamicton earlier described in western Estonia (Kadastik & Kalm 1998). Genesis of this unit remains unclear. According to varve correlation between the studied sequences an interval of 20 silty varves with dispersed sand grains and few dropstones stratigraphically corresponds to the aforementioned unit D. At the lower altitude of the clay distribution area the upper dehydrated clay complex (unit A) as well as the complex of very thin distal varves (unit B) is missing. This together with the erosional slope in clay surface in the northern part of discussed area, are interpreted as remarkable water level drop accompanied by wave erosion at the end of Late Weichselian

A 360 yrs long local varve chronology for the Pärnu area has been established through the correlation of 18 varved sequences (Fig. 2). The correlation has been done through the straight comparison of sediment sequences (Fig. 1) being thus more convincing than classical varve graph correlation. All studied sequences contain easily recognizable marker interval of 20 silty varves with a colour change from light grey to reddish–brown on top of the interval, which certainly strengthen the correlation. One sequence from the central part of the Pärnu Bay contained 320 varves of only distal and normal (formed in the central part of the proglacial basin) varve series. In that site, we could not penetrate the whole varve series and did not reach the mentioned marker interval. Thus, so far this sequence can only tentatively be correlated to the chronology, but if the presented correlation to be true, the chronology for Pärnu area will extend to ca 460 yrs.

All studied varve sequences display a normal varve series reflecting a gradual ice recession within at least 360 yrs and corresponding change in sedimentary conditions in the proglacial lake where, at each point, proximal conditions gradually changed to conditions that are more distal. There is no evidence from the varved clay studies pointing to the stagnation or readvance of the ice terminus prior or during the formation of the Pandivere–Neva ice marginal formations. As derived from the beginning of varve formation in each studied site the ice recession rate from 95–160 m/yr could be calculated prior to the Pandivere–Neva stade, which is in accordance with the earlier studies (Hang 2001, Kalm 2006) from Estonia and adjoining areas (Saarnisto & Saarinen 2001). Further studies are needed to interpret the widely reported marker interval with 20 silty varves and with the sharp colour change on top of the interval.

Fig 1. An example of varved clays from Pärnu area, western Estonia. Arrow points to the marker level with rapid colour change from light grey to reddish–brown on top of the series of 20 silty varves containing dispersed grains of sand. Right figure demonstrates a straight comparison of different varved sequences for varve correlation.

Landslides in proglacial clays

Environmental hazards connected with the proglacial varved clays are attributed to the landslides. Usually landslide hazard is a rare occurrence on low laying (<20 m a.s.l.) coastal plains with slightly undulating topography like the surroundings of the Pärnu City. However, in recent years 9 sliding events out of 25 historically recorded landslides occurred at the rive valleys cut into the proglacial varved clay. The river valleys are typically 10…15 m deep. Steepest slopes (25–30º) and strongest erosion occur at the bends of river meanders. A principle geological setting displays the bedrock surface from Silurian limestones and Devonian sandstones at an altitude of –10 – –15 m and it is covered by grey loamy till of the Late-Weichselian age. The latter is overlain by 5–9 m (max 30 m) thick annually laminated glaciolacustrine silt and clay. Thickness of the Holocene marine sand and silt, covering the glaciolacustrine deposits is normally 2–3 m and as a maximum 10 m. Geotechnically weakest soil type in the study area is glaciolacustrine varved clay.

In the clay fraction (<0.002 mm) of the sediment illite (70–80%), smectite-illite (9–19%), kaolinite (6–8%) and chlorite (2–5%) dominate, while the silty summer layers are composed primarily by quartz (46–59%), feldspars (14–18%), carbonate (11–21%) and in minor amounts

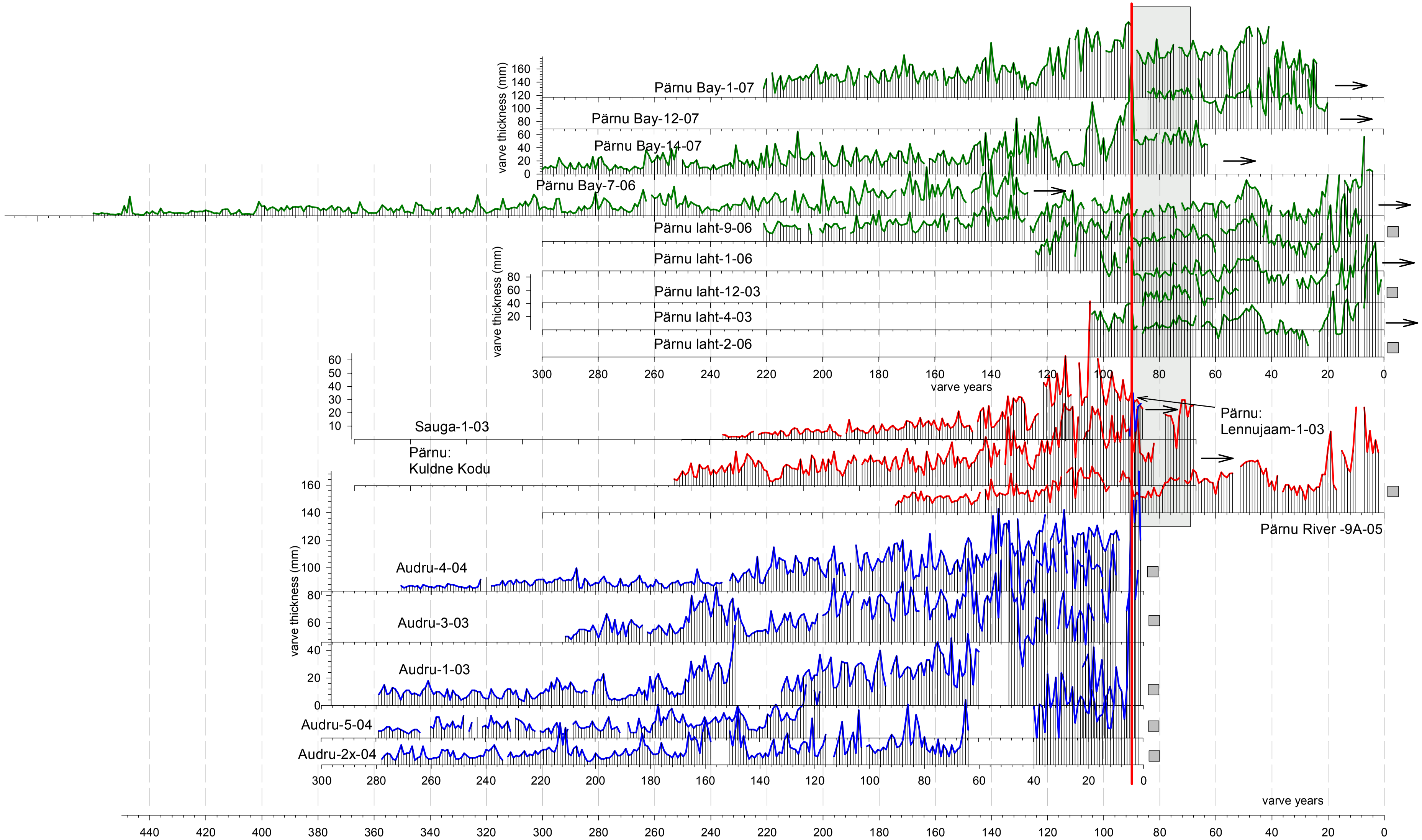

Fig. 2. A local varve chronology for the Pärnu proglacial basin of varved clays in western Estonia. Green graphs display the sequences obtained from the Pärnu Bay eastern Baltic Sea; red colored graphs display the sequences from the City of Pärnu and blue color marks the sequences obtained from the Audru site at the distal slope of the Pandivere-Neva ice marginal formations c 10 km from Pärnu. Shadowed area and the red line mark the marker interval with the color change reported in all connected sequences. Arrow points that the bottom varve has not been reached and the square points that the bottom varve has been reached.

by other minerals. Geotechnical parameters of clays in the area change vertically while conversion is even and without rapid changes. According to Casagrande plasticity chart (Coduto 1998), the upper portion of these clays can be classified as CL (fat clay) and lower portion as CH (lean clay). Water content of the clay decreases towards the depth being 70–90% in the upper portion and decreasing downwards to 30–60%

Two groundwater aquifers inside the Quaternary sediments have been reported in the Pärnu area. The upper layer, unconfined, is bound to marine sands, and the lower, confined one, to the till under the glaciolacustrine clay. Glaciolacustrine clay acts as an aquiclude between the two groundwater aquifers. Groundwater level in surface sediments is changing according to weather conditions: during droughts, there may be no water at all in sediments above the clay while during thaw or heavy rains the water level rises up to –1 m from the surface. Second aquifer related to the till is pressurized and the piezometric level of this aquifer has been between 0...1.5 m a.s.l. in the city of Pärnu over the last 7 years (Kohv et al. 2009 in press).

Nine recent landslides were studied in detail: their geometry was instrumentally measured, sediment description and analysis was done based on core or outcrop material. The depth and configuration of the sliding surface was measured from a number of corings. Annual lamination in the whole package of the glaciolacustrine clay with distinct summer and winter layers allowed the tracing of landslide-deformed lamina throughout the whole cross section of a slide. Stability of slopes eroded into the glaciolacustrine deposits and the slope failures at the sites of the measured landslides were evaluated using modelling with Janbu corrected method. Hydrogeological conditions were also incorporated to the model, for example the groundwater level 1 meter below the surface is reflecting the high water-stand conditions. Additional shear stress, which is generated by the groundwater flow in the sandy slope, was calculated using finite element method, which is incorporated to the slope stability software Slide v 5.0 (Rockscience Inc.).

Three different groups (A, B, C) of landslides have been distinguished based on soil type involved, sediment stratigraphy and failure mechanism:

A- The largest slides (up to 80 x 137 m) occurred in the glaciolacustrine clayey soils on the erosional riverbanks or on the straight sections of the valleys. Their rupture surface penetrated the whole varved clay section and the sliding material always reached the river channel pointing to the importance of fluvial erosion and corresponding channel morphology to promote slope failure. These were retrogressive complex slides. Critical slope angle in natural conditions for the group A landslides is $>10^{o}$.

B- Slides in the marine fine–grained sand appeared in the upper parts of the valley slopes, the rupture surface went through the marine sand and in some cases through the topmost part of the varved clay. The sliding body never reached the channel. The slides were triggered by extra shear stress generated by groundwater movement during the high stand of the ground water level. Critical slope angle in natural conditions for the this group is $>20^{o}$

C- Small landslides in varved clay were observed only on the banks of channels, occurred in groups mainly on the erosional bank and their rupture surface went through the upper 2…3 m section of the varved clay. They have been generated by the river erosion and triggered by the rapid waterlevel changes. Group C slides are first stages of the larger, group A slides.

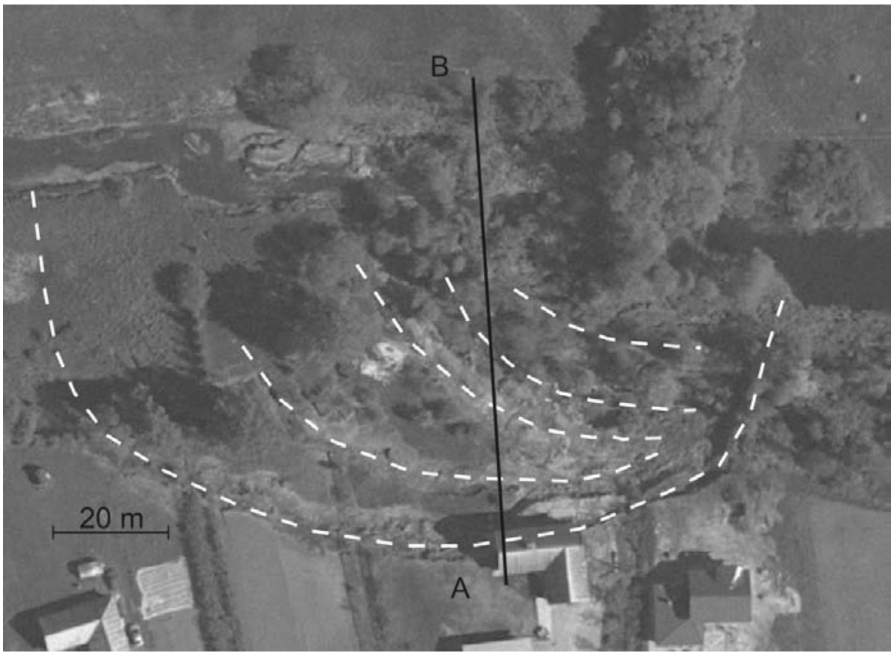

Fig. 3. Sauga landslide is one of the large complex landslides in varved clay (group A). White dashed lines mark the scarp and the transverse ridges. Black line marks the location of the geological profile on Fig. 4.

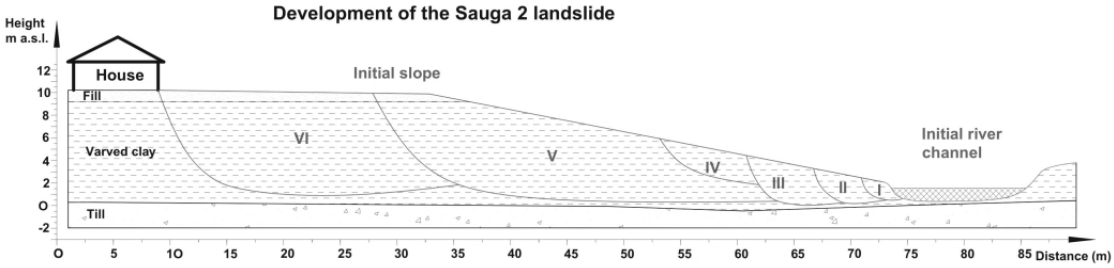

Fig 4. Profile of the Sauga complex retrogressive landslide. Landslide stages from slope stability modelling are shown by Roman numbers.

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International Field Symposium
September 14–17, 2009

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