Geological Quarterly, 2004, **48** (1): 15–22

Glaciotectonic deformation patterns in Estonia

Maris RATTAS and Volli KALM

Rattas M. and Kalm V. (2004) — Glaciotectonic deformation patterns in Estonia. Geol. Quart., **48** (1): 15–22. Warszawa.

Field and archival data have been used in compiling the Glaciotectonic Map of Estonia. Two principal types of glaciotectonic deformations shown on this are discussed here: dislocations of rigid bedrock, and soft bed deformations associated with unconsolidated drift masses. Most bedrock disturbances occur in the narrow zone south of the Baltic Klint and in the tectonically crushed zone where the fractured bedrock was easy to break, displace and deform by the moving glacier. Some of the bedrock dislocations are related to ice-marginal deposits of the Late Weichselian Glaciation (Palivere and Pandivere Phases). Most subglacial deformations of soft sediments are simple in style, namely: shear and ductile deformations within a thin layer. The spatial organisation and efficiency of drainage beneath the local ice streams determined the deformational behaviour of sediments at the ice/bed interface. Ice-marginal deposits of the Late Weichselian deglaciation have not been subjected to large-scale compressive deformation. This suggests that most marginal deposits were formed as the result of brief standstills of the ice margin which caused sediment deformation either at the ice margin or beneath the ice sole.

Maris Rattas and Volli Kalm, Institute of Geology, University of Tartu, Vanemuise 46, EE-51014 Tartu, Estonia; e-mails: mrattas@ut.ee, volli.kalm@ut.ee *(received: January 9, 2003; accepted: October 28, 2003).*

Key words: Estonia, Quaternary, glaciotectonic landforms and structures, bedrock dislocations, soft sediment deformations, drumlins.

INTRODUCTION

Unconsolidated deposits and bedrock which have been dislocated by ice masses are well-known features in formerly glaciated areas. These deformed sequences have been interpreted as of glacial origin and they profoundly influence the stratigraphical and lithological variability of Quaternary deposits and the shallow subsurface of pre-Quaternary rocks (Aber, 1982). The presence or absence of glaciotectonic features is related to the nature of both ice and substrate, which vary temporally during a single glaciation or during successive glaciations (Aber *et al*., 1995). Therefore, they can help characterise the evolution of both the ice masses and the palaeogeography of the regions where they occur.

During the Pleistocene glaciations Estonia was affected by the Baltic and Peribaltic ice streams which moved mainly to the south and south-east (Raukas and Gaigalas, 1993). The flow of the local ice streams was controlled by the underlying bedrock topography beneath the ice sheet. Some ice flows were restricted to areas covered by soft sediments, where their motion was controlled by fluctuations in water pressure which provided effective lubrication for basal sliding and subglacial till deformation (Rattas and Piotrowski, 2003).

Northern and western Estonia have been the areas of strong erosion and the Pleistocene cover is usually less than 5 m thick. Southern Estonia has been an area of predominant deposition and the Pleistocene deposits are typically 50–200 m thick. This uneven distribution of the Pleistocene cover means that bedrock deformations are predominantly observed in the northern and western parts of Estonia, while soft-sediment deformations occur mainly in the south (Fig. 1).

The first glaciotectonic review map of Estonia has been compiled as a part of the Central European Glaciotectonic Database Project (CEGDP). In this, glaciotectonic features in the Estonia have been classified into five main categories (Rattas and Kalm, 1999) (Fig. 1):

1. Ridges and hills built of disturbed Quaternary sediments with a clear topographical expression. These glaciotectonic landforms (push moraines, hummocky moraines, cupola hills, drumlins and eskers) mostly denote ice-marginal positions of a retreating ice sheet. A system of such individual glaciotectonic landforms, which comprises largely unconsolidated Quaternary strata, is termed a composite massif. The most expressive composite massifs are the Otepää and Haanja heights, and the Karula and Sakala uplands.

2. Point occurrences of buried disturbed sediments (in individual boreholes or exposures), not expressed in the present

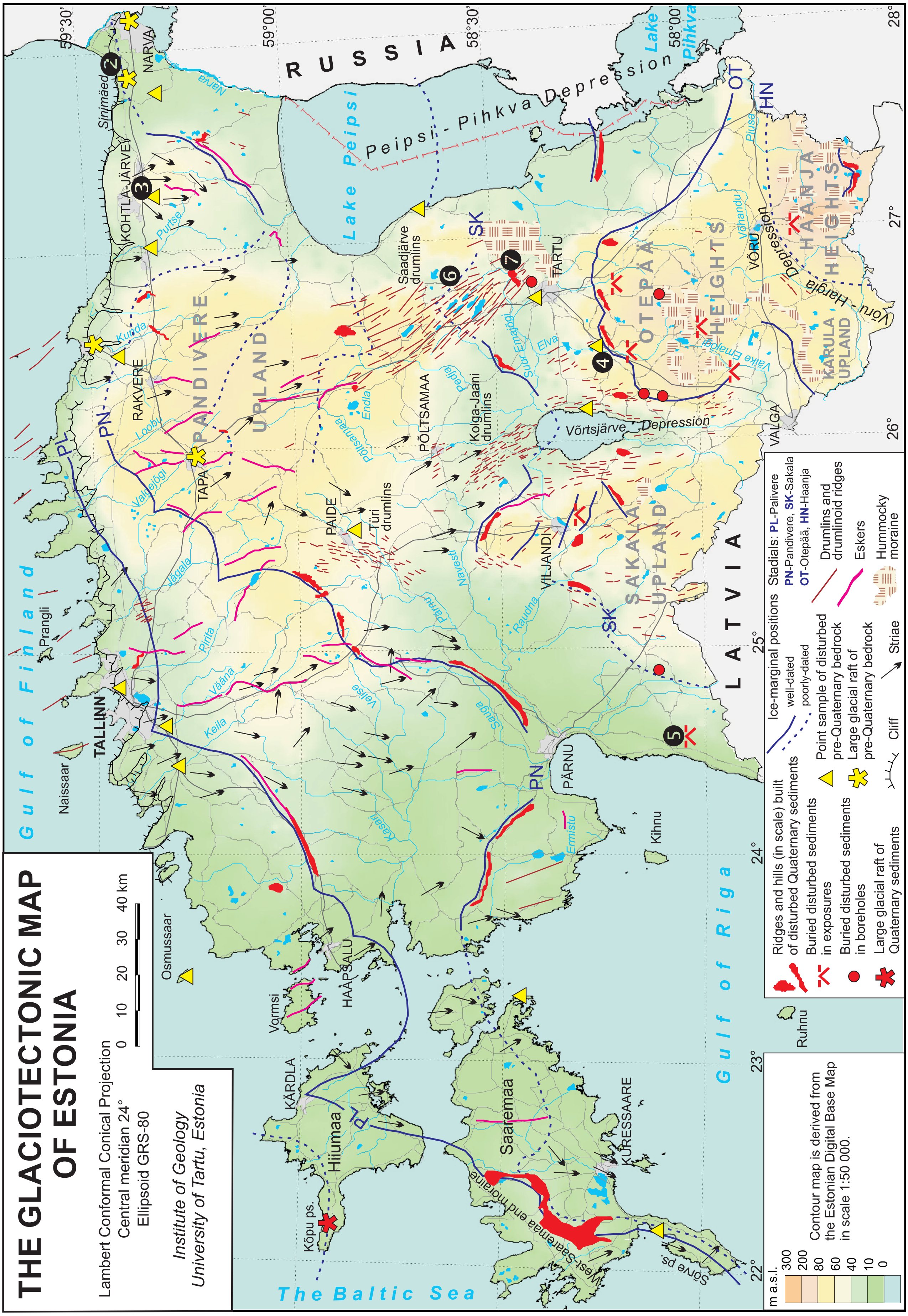

Fig. 1. Glaciotectonic Map of Estonia

Bedrock dislocation is shown in yellow, soft-sediment deformation in red; locations of the sites discussed in the text are shown as black circles with numbers of relevant text-figures

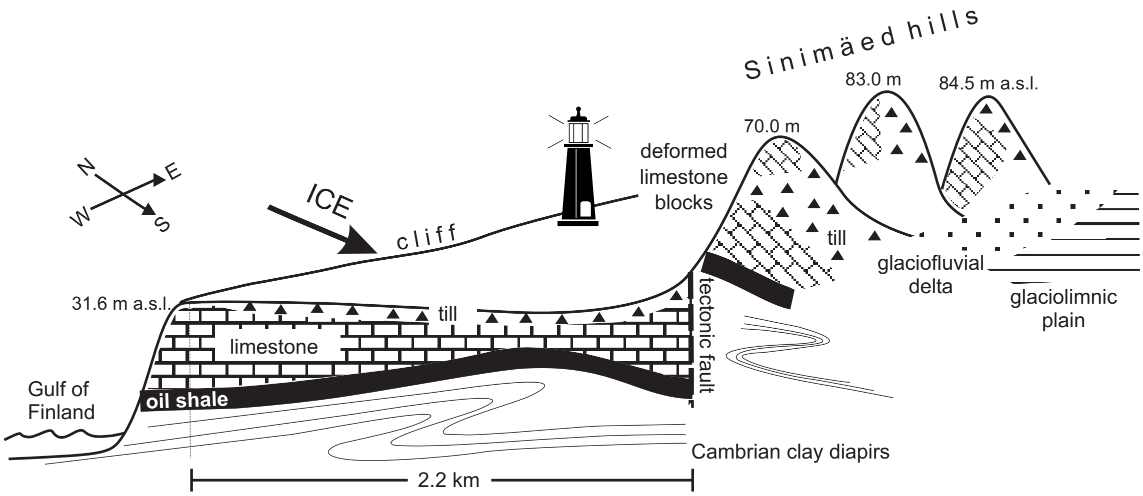

Fig. 2. Schematic section of the Sinimäed hills (*cf.* Jaansoon-Orviku, 1926; Miidel *et al.*, 1969)

For location see Figure 1

landscape. Most interglacial (Holsteinian and Eemian) strata in Estonia are dislocated and form erratics in glacial deposits.

3. Point occurrences of disturbed pre-Quaternary bedrock with no morphological expression. Hard bedrock dislocations, such as fractures, folds, and fissures have been recognised in exposures and borehole records.

4. Large glacial rafts, expressed in the present landscape, are allochthonous masses of dislocated bedrock or Quaternary deposits. These have been dragged or pushed to the surface; some are partly or entirely buried under or within thick Quaternary successions.

5. Large glacial depressions, which have been subjected to intense glacial erosion during several glaciations are reflected in the present-day landscape and bedrock topography (Lake Peipsi and Võrtsjärve depressions). A few bedrock deformations initiated by lateral glacier pressure have been observed near the outlines of depressions. Soft sediment deformations have been suggested in the Baltic Sea from seismic data (Noormets and Flodén, 2002).

Most of the glaciotectonic structures observed in Estonia were formed during the last, Late Weichselian Glaciation. However, some concealed deformation complexes, such as dislocated interglacial deposits and bedrock blocks, may have been formed during earlier glaciations.

In this paper we describe the glaciotectonic map of Estonia, which was compiled for the CEGDP. Data presented on this map have been derived from various sources, former reviewed in Rattas and Kalm (1999). Many of the glaciotectonic features have only been noted as evidence of the active ice; only a few have been described in detail (Jaansoon-Orviku, 1926; Orviku, 1930, 1936; Miidel *et al*., 1969; Heinsalu, 1970; Levkov and Liivrand, 1988; Karukäpp *et al*., 1996). Detail investigations have been carried out to study the subglacial deforming bed processes which formed the drumlins (Rattas and Kalm, 2001*a*; Rattas and Piotrowski, 2003).

In addition, we describe glaciotectonic structures in Estonia strongly influenced by substrate type: a — dislocations of rigid bedrock and b — deformations of unconsolidated drift masses. These different deformation styles are related to local ice streams dynamics during the Late Weichselian Glaciation.

BEDROCK DISLOCATIONS

The pre-Quaternary bedrock in Estonia includes Ordovician and Silurian limestones and dolomites in northern Estonia and poorly consolidated Devonian siltstones and sandstones in the south. To the north of the limestone belt and the Baltic Klint (see cliff Fig. 1) a narrow band of the Cambrian sandstone and unconsolidated clay crop out.

The pre-Quaternary surface has been highly glaciotectonised. Folds, fractures, fissures overthrusts and erratic megablocks of rigid bedrock have been observed in several exposures, open-cast quarries, and boreholes. Most bedrock dislocations have formed *in situ,* only few erratic megablocks having been moved for a short distance (2–3 km) from source.

Major glaciotectonic dislocations within carbonate rocks are concentrated in the narrow zone to the south of the Baltic Klint (Fig. 1). This, relatively high (up to 50–60 m) hardrock structure in the path of southerly moving ice caused changes in glacier-bed dynamics and subglacial strain conditions. Most deformation has been located in a tectonic crush zone, where the solid bedrock has been divided into a number of blocks of different sizes by Palaeozoic and Neogene-Quaternary crustal movements (Miidel and Vaher, 1997; Puura and Vaher, 1997). Fractured bedrock was relatively easy to break, displace or deform by the moving glacier.

Large glacial rafts of pre-Quaternary bedrock, transported for some kilometres away from the cliff, are located at Sinimäe in NE Estonia (Fig. 1). Three bedrock blocks of Lower Ordovician limestone form the cores of three hills at Sinimäed, forming an E–W oriented ridge 5 km long. The ridge rises 40–50 m above the generally flat topography, about 2.2 km south of the Baltic Klint (cliff) edge (Fig. 2). The limestone core crops out on the northern and western slopes of the hills. The limestone beds are strongly inclined (18–70°), folded and locally

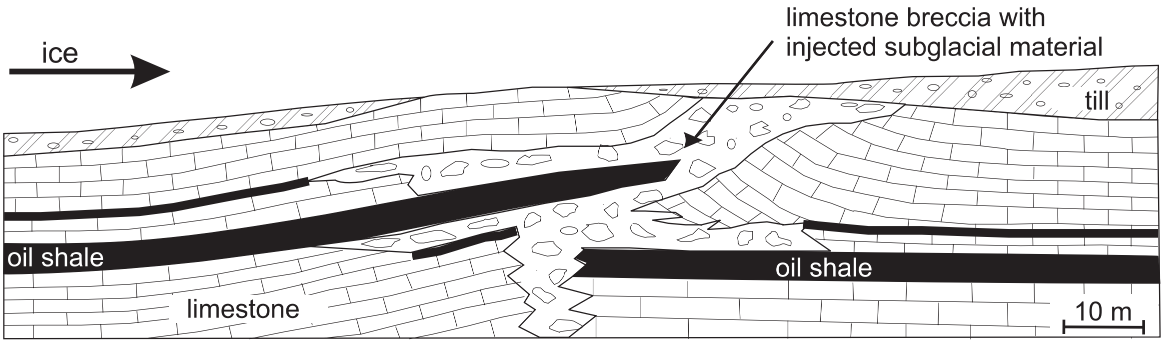

Fig. 3. Concealed structure (overthrust) in hard bedrock induced by ice stress (modified from Miidel *et al.*, 1969 and Heinsalu, 1970)

For location see Figure 1

overthrusted. The presence of till lenses intercalated within the limestone beds and folds indicate the glaciotectonic origin of this dislocation (Miidel *et al.*, 1969). Consequently, it has been interpreted as a glacial marginal complex (push moraine) including deformed bedrock blocks proximally and a glaciofluvial delta distally (Raukas *et al.*, 1971). The bedrock blocks may have been pushed in front of the advancing glacier or dragged along subglacially. The Sinimäed hills mark the ice margin of the Pandivere Phase (12.480–12.230 ^{14}C years BP, Raukas *et al.*, 2004) of the Late Weichselian ice advance.

Orviku (1960) proposed that the limestone blocks of the Sinimäed hills were squeezed upwards by Cambrian clay diapirs along pre-existing tectonic faults. Similarly, in the north-easternmost part of Estonia, between Sinimäe and Narva, some of the narrow northerly and northeasterly oriented linear bedrock structures have been related to diapirs of Cambrian clays (Vaher and Mardla, 1969). Unconsolidated clay has been squeezed upwards along zones of minimum resistance where the clay thickness has been tripled (Puura and Vaher, 1997). It has been suggested that the diapirs might have formed along pre-existing disturbances (Puura and Vaher, 1997). However, the glacier ice load and dynamic pressure during ice advances may also have facilitated clay deformation.

Several bedrock disturbances in the northeastern Estonia have been recorded from boreholes. The glaciotectonic origin of such concealed faults and overthrusts is, though poorly constrained. Fractured bedrock has usually been displaced by shearing in response to the injection of subglacial material into karst cavities or sub-horizontal fractures, and subglacial till usually contains plucked or loosened bedrock fragments (Figs. 1 and 3).

Poorly consolidated Devonian terrigenous bedrock (sandstone and siltstone) has mostly been folded and fractured by subglacial deformation. A thin deformed layer occurs at the till/sandstone contact zone. The deformed layer shows either small shear structures or diffuse mixing of till and sandstone with till rafts or sandstone blocks (Fig. 4).

DEFORMATION OF SOFT SEDIMENTS

A number of glaciotectonic landforms (composite massifs, ridges, cupola hills, drumlins, glacial depressions) composed of glacially deformed soft sediment material have been recognised (Fig. 1; *cf.* Rattas and Kalm, 1999). Buried disturbed sediments, not expressed in the present landscape, have been also documented.

The pattern of regional Pleistocene stratigraphy (Raukas and Kajak, 1995) suggests that the original bedding of most interglacial deposits in Estonia is disturbed and they occur as erratics in glacial deposits (Liivrand, 1991).

The Haanja and Otepää heights, and the Karula Upland in southern Estonia are the best-expressed composite massifs, built of heaped and subsequently deformed drift masses. These composite massifs are characterised by the great thickness of the component glacial deposits and by a very complicated relief

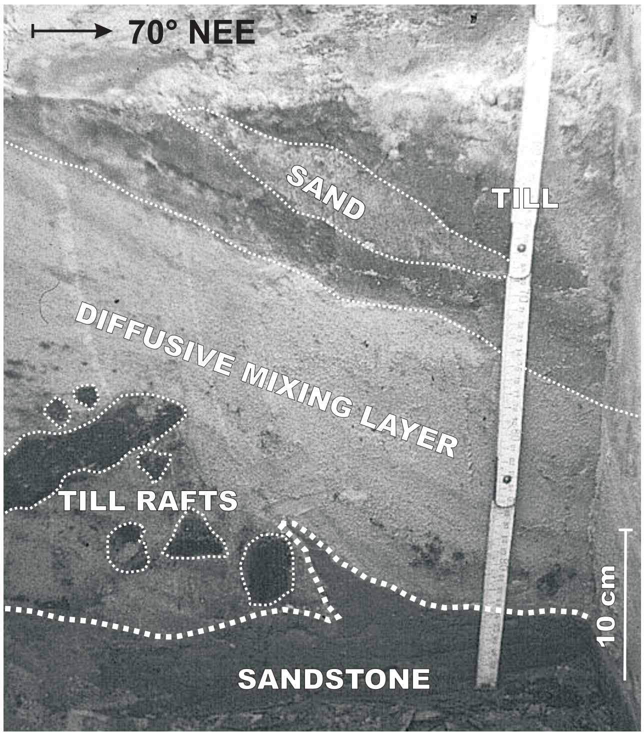

Fig. 4. Glacially deformed layer in the contact zone between poorly consolidated sandstone and glacial deposits

For location see Figure 1

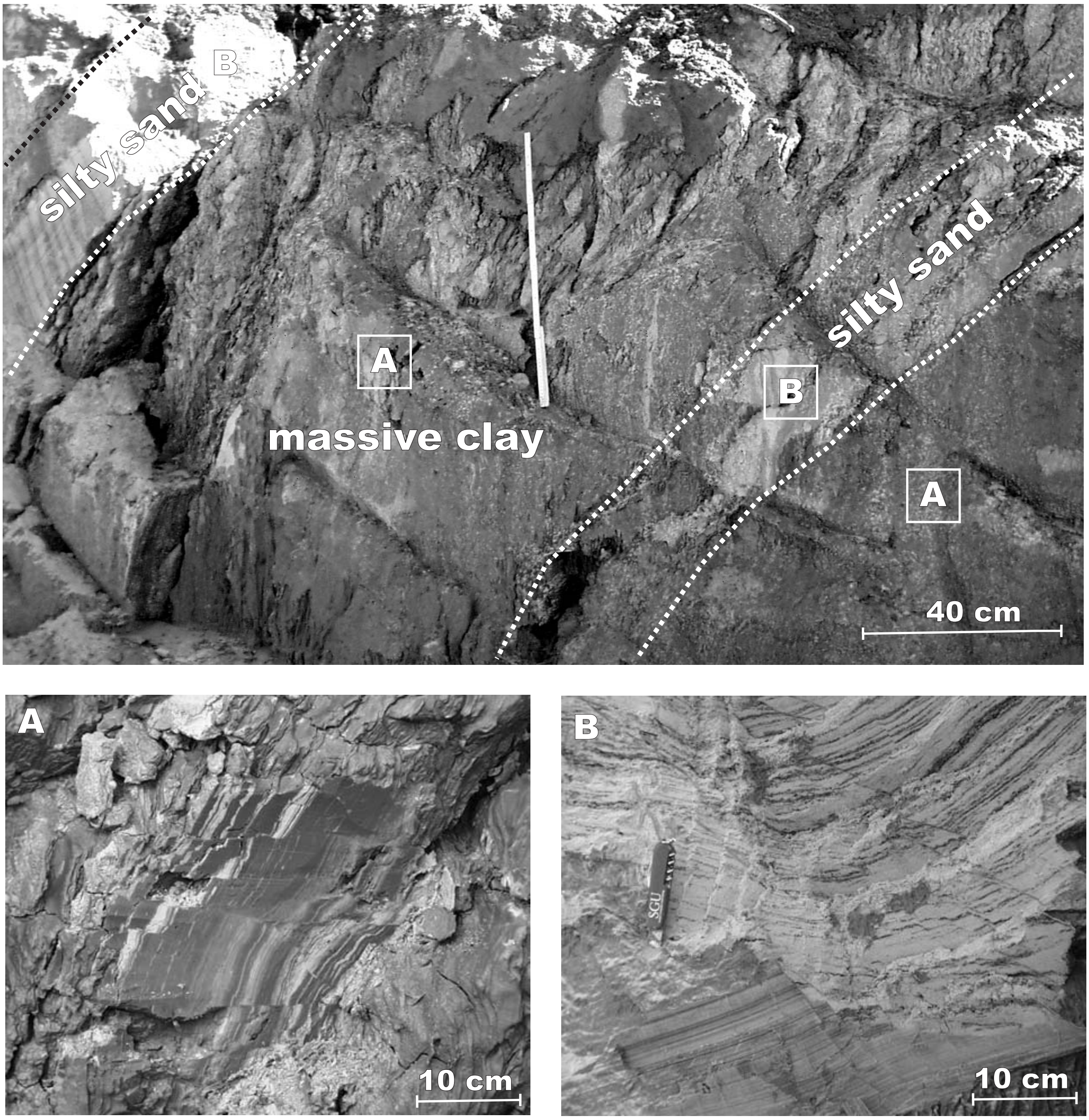

Fig. 5. Steeply tilted silty sand layers in massive clay

Small faults and shear planes occur in finely laminated silty sand — A as well as in massive clay —B (Arumetsa clay quarry, north-west Estonia); for location see Figure 1

arising from stagnant ice (kames, morainic hummocks). However, a few forms reflecting active glacier ice (drumlins, end moraines) occur within the limits of these composite massifs and these contain glaciotectonic structures internally. These massifs probably also contain glaciotectonic deformations associated with stagnant ice, i.e. ice collapse and crevasse infill structures; such features of glaciokarst have been discussed by Raukas and Karukäpp (1994).

The ice-marginal positions of different stadials or oscillations of the last, Late Weichselian deglaciation at about 13 500–11 000 ^{14}C BP (Pirrus and Raukas, 1996; Raukas *et al.*, 2004) are marked by discontinuous chains of push moraines, marginal eskers, and hummocky moraines, consisting of either disturbed diamicton or waterlain sediments. These ice shoved ridges and hills contain listric thrusts, faults, and folds associated with proglacial compressive deformation, or of folding and shearing beneath moving ice. As Hart and Boulton (1991) have argued, compressive (proglacial) and extensional (subglacial) deformations are related to one another, and are superimposed upon one another as the ice sheet advances and retreats.

A substantial deformed succession has been observed in the Arumetsa clay pit in the southwestern Estonia, where the uppermost 4 m of clays and laminated silts and sands are strongly deformed. The orientations of folds and fractures suggest the glacier push was from the north (Fig. 5). The deformed sequence is covered by a 0.5–1 m thick sandy till layer, the boundary be-

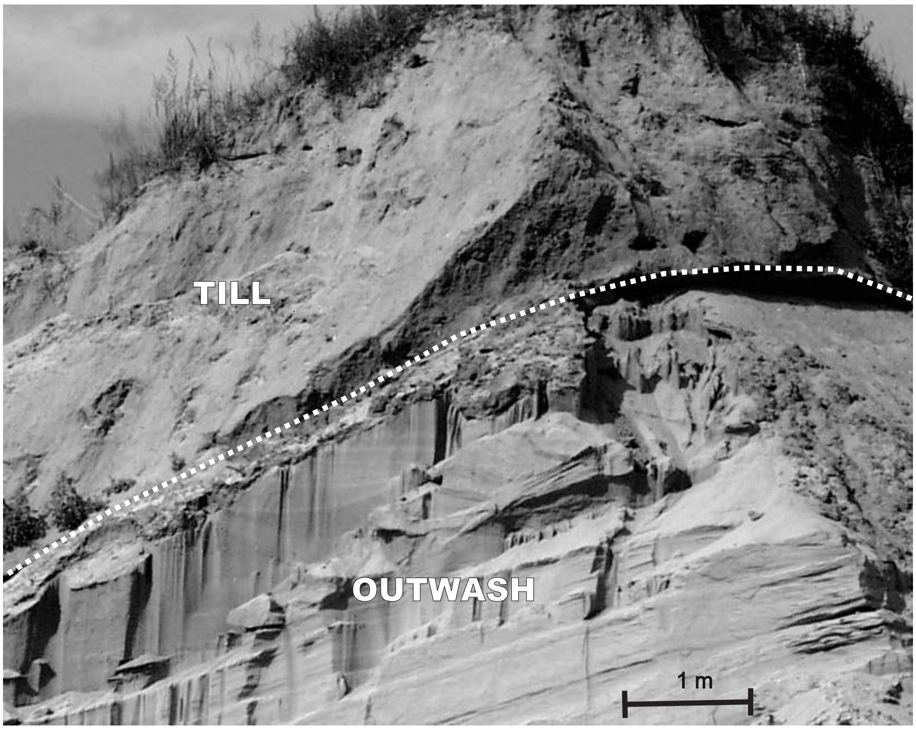

Fig. 6. Sharp basal contact of till overlying undisturbed outwash sediments, Saadjärve drumlin field

For location see Figure 1

tween till and deformed strata not being clearly observable. As the deformed sequence is not expressed in the present topography, most likely it has been deformed subglacially during the last deglaciation or even during previous glaciations.

Deformation of soft subglacial sediments has been invoked to account for the origin of drumlins (Smalley and Unwin, 1968; Boulton, 1987; Smalley and Piotrowski, 1987; Hart, 1997; Hindmarsh, 1998). In Estonia the drumlins are mostly built of soft deposits of Pleistocene age, and only occasionally have a bedrock core (Rõuk and Raukas, 1989). Studies of the internal structures of drumlins have been concentrated on only a few drumlins because good exposures are rare. More detailed investigations on the subglacial deforming bed processes on the formation of drumlins have been carried out in the Saadjärve drumlin field in eastern Estonia (Fig. 1). The morphology of the Saadjärve drumlins is given by the uppermost massive sandy till which was formed during the drumlinization, initiated by the Late Weichselian ice sheet re-advance (between the Otepää and Pandivere Phases 12 600–12 050 ^{14}C BP). Beneath the drumlin-forming till cover the drumlins showed mostly an undisturbed core of older tills and outwash sediments (Rattas and Kalm, 2001*b*; Rattas and Piotrowski, 2003). Drumlins are the result of net subglacial deforming bed erosion and deposition with evidence of varying degrees of deformation. In a few drumlins there is no visible evidence of deformation or of a subglacial deforming bed. Here, there was either no bed deformation or this may have been restricted to a very thin (mm-scale) zone. In the proximal part of the field, where the till rests on outwash sediments, the till/outwash boundary is typically gradational with a several-cm-thick transition zone enriched in material derived from the underlying meltwater complex. In some cases, the boundary is very sharp with no diffusive mixing and no vertical changes in till texture and structure were seen (Fig. 6). A visibly deformed bed has been observed in the distal part of the drumlin field (Rattas and Kalm, 2001*a*), where the drumlin-forming till often contains large intensely contorted lenses and pockets of outwash sand in its lower part. Contacts between sand pods and the surrounding till matrix are smudged and gradational indicating mixing and granular sediment diffusion. Figure 7 depicts large-scale ice-induced shear structures on the flank of a drumlin. The drumlin-forming till is folded in its lower part and a large sand lens has been thrusted into the till. The till contains also thin subhorizontal layers and lenses of waterlain deposits. These could be interpreted as drag features or washout features formed during basal decoupling. These observations suggest ductile deformation, when the drumlin-forming glacier was moving on a weak soft till or on outwash sediments (Rattas and Piotrowski, 2003).

DISCUSSION AND CONCLUSIONS

Glaciotectonic processes have exerted a remarkable influence on the structure of the upper part of the bedrock and on Quaternary strata in Estonia. The Glaciotectonic Map of Estonia represents a synthesis of current knowledge concerning the types and locations of glaciotectonic landforms and structures in Estonia. Many more examples of such deformation are likely to exist. The regional stratigraphy suggests that most of the glaciotectonic structures were formed during the last, Late Weichselian Glaciation by Baltic and Peribaltic ice streams.

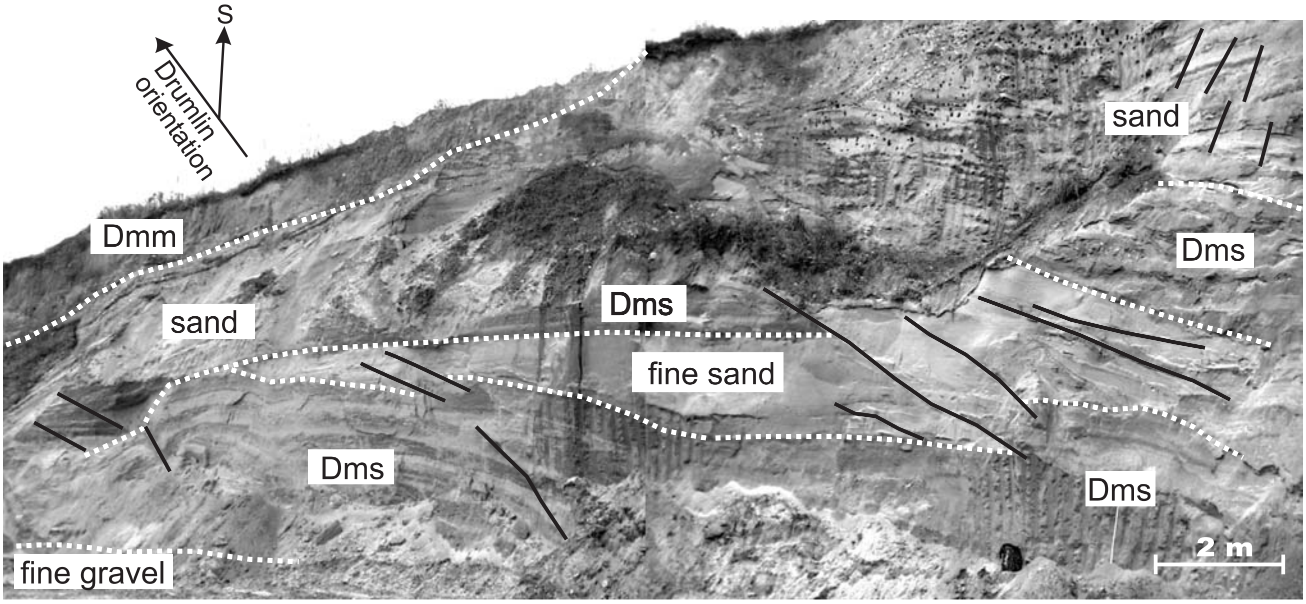

Fig. 7. Deformed subglacial sediments in drumlin due to ice-induced shear (Saadjärve drumlin field)

Thrust and related displacements are accentuated by solid lines, lithological boundary by a dashed line; the drumlin-forming till: Dmm — massive matrix-supported diamicton, Dms — massive stratified diamicton; for location see Figure 1

However, some concealed deformation complexes, such dislocated interglacial deposits and bedrock dislocations, may have been formed during earlier glaciations.

Two types of glaciotectonic deformations associated with different substrate lithologies, i.e. hard bedrock dislocations and soft bed deformations, have been described in this paper. The active movement of ice produces both. The distribution of both types is primarily related to the presence/absence and thickness of the Pleistocene cover. Dead ice structures associated with melting ice-blocks have not been not studied in detail. Most of hard bedrock disturbances occur in the narrow zone to the south of the Baltic Klint. Tectonically fractured Palaeozoic bedrock was easy broken, displaced and deformed by the moving glacier. Some of the bedrock dislocations may be related to the ice-marginal zones of the Late Weichselian ice re-advance (Palivere and Pandivere Phases) situating along the marked ice margin lines. The glacial origin of some concealed bedrock disturbances is poorly constrained, but the bedrock faults and fractures locally contain injected subglacial material and plucked bedrock.

Glaciotectonically folded sequences with scattered faults in poorly consolidated silt- and sandstone have a subglacial origin, a thin deformed bed occurring in the bedrock/till contact zone, which is typically transitional and smudged. In some cases, on Devonian bedrock, these deformations can be classified also as subglacial soft bed deformations, where the uppermost part of the Devonian consists of unconsolidated sand or silty sand deposits.

Soft sediment deformations were formed either proglacially by compressive deformation, represented by push moraines, or by subglacial deformation by shearing. In Estonia the ice-marginal deposits have not been subjected to large-scale compressive deformation. This indicates that most of the ice--marginal deposits were formed as a result of standstills of the ice margin, being only for short time intervals in an equilibrium state, caused sediment deformation either at the ice margin or beneath the ice sole. The re-advance character of the Palivere Phase is clearly identified by the distribution of push moraines and by the sedimentary record (Raukas *et al.*, 1971), but no large deformation has taken place. The lack of deformations can be explained by the scarcity of (pro)glacial deposits in northern and northwestern Estonia, the area being subjected to erosion — dominated processes throughout the Pleistocene.

Most examples of subglacial deformation of soft sediments show a very simple style of deformation: shear and ductile deformations, which are restricted to a thin deformed layer, from a few centimetres to one metres below the glacier sole. Only in Arumetsa has a 4 m thick deformed layer of soft sediment been observed. The deformation history of these deposits is unclear and complicated. It is likely that the deformation took place during earlier glaciations by ice surges or by compressive deformation. In this case the deformed layer was discordantly covered by Late Weichselian till.

The example of the Saadjärve drumlin field indicates that fast ice flow was initiated in the southeastern part of the field, where subglacial pore-water pressure builtup was greatest and the thickest deformed layer is observed (Rattas and Piotrowski, 2003). The drumlin field indicates an onset area of the local ice stream, which was funneled into a narrow track by the larger Peipsi and Central Estonian ice lobes. The measurements of structural features have showed a dominant stress direction from the west (Rattas and Kalm, 2001*a*).

The distribution of glaciotectonic features in Estonia is constrained by the presence of a thin deformable substratum. Deformation studies emphasize how the spatial organisation and efficiency of drainage beneath local ice streams determined the deformational behaviour of sediments at the ice/bed interface.

Acknowledgements. We are deeply indebted to Dr. A. Ber (Polish Geological Institute, Warsaw) for valuable discussions during our joint activities under the Central European Glaciotectonic Database Project of the INQUA Commission on Glaciation. Prof. Leszek Marks (Polish Geological Institute, Warsaw) and Prof. Hanna Ruszczyńska-Szenajch (University of Warsaw) are thanked for their careful reviews that led to significant improvements to this paper. Jan Zalasiewicz is thanked for improvement of the English language. The research was financially supported by the Basic Funding grant no.0182530s03 from the Estonian Ministry of Education and Research.

REFERENCES

ABER J. S. (1982) — Model for glaciotectonism. Bull. Geol. Soc. Denmark, **30**: 79–90.

ABER J. S., BLUEMLE J. P., BRIGHAM-GRETTE J., DREDGE L. A., SAUCHYN D. J. and ACKERMAN D. L. (1995) — Glaciotectonic Map of North America, 1:6 500 000. Geol. Soc. Am. Maps Chart. Ser., MCH 079: 1–7.

BOULTON J. S. (1987) — A theory of drumlin formation by subglacial sediment deformation. In: Drumlin Symposium (eds. J. Menzies and J. Rose): 25–80. Balkema, Rotterdam.

HART J. K. (1997) — The relationship between drumlins and other forms of subglacial glaciotectonic deformation. Quat. Sc. Rev., **16**: 93–107.

HART J. K. and BOULTON G. S. (1991) — The interpretation of glaciotectonic and glaciodepositional processes within the glacial environment. Quat. Sc. Rev., **10**: 335–350.

HEINSALU Ü. (1970) — Aluspõhja glatsiotektoonilisi lasumusrikkeid Kirde-Eestis. Eesti Loodus, **2**: 118–120.

HINDMARSH R. C. A. (1998) — Drumlinization and drumlin-forming instabilities: viscous till mechanisms. J. Glaciol., **44**: 293–314.

JAANSOON-ORVIKU K. (1926) — Über die Glazialschollen in Eesti (in Estonian with German summary). TÜ Loodusuurijate Seltsi Aruanded, XXXIII, **1**: 48–56.

KARUKÄPP R., MOORA T. and PIRRUS R. (1996) — Geological events determining the Stone Age environment of Kunda. In: Coastal Estonia: Recent Advances in Environmental and Cultural History (eds. T. Hackens *et al.*). Rixensart, Belgium, PACT, **51**: 219–229.

LEVKOV E. and LIIVRAND E. (1988) — On glaciotectonical dislocations of interglacial deposits in Karuküla and Kõrveküla sections (Estonia) (in Russian with English summary). Proc. Acad. Sc. Est. SSR. Geology, **37** (4): 161–167.

LIIVRAND E. (1991) — Biostratigraphy of the Pleistocene deposits in Estonia and correlations in the Baltic region. Stockholm University. Dept. Quat. Res., Report 19, Doctoral Thesis.

MIIDEL A. and VAHER R. (1997) — Neotectonics and recent crustal movements. In: Geology and Mineral Resources of Estonia (eds. A. Raukas and A. Teedumäe): 177–180. Est. Acad. Publishers, Tallinn.

MIIDEL A., PAAP Ü., RAUKAS A. and RÄHNI E. (1969) — On the origin of the Vaivara Hills (Sinimäed) in NE Estonia (in Russian with English summary). Proc. Acad. Sc. Estonia. Chem., Geol., **18** (4): 370–376.

NOORMETS R. and FLODÉN T. (2002) — Glacial deposits and Late Weichselian ice-sheet dynamics in the northeastern Baltic Sea. Boreas, **31**: 36–56.

ORVIKU K. (1930) — Die Glazialschollen von Kunda-Lammasmägi und Narva-Kalmistu (Eesti). Sitzungsberichte der Naturforscher-Gesellschaft bei der Univ. Tartu, **34** (3, 4): 174–179.

ORVIKU K. (1936) — Kihitussiirdeid Eesti aluspõhjas. Eesti Loodus, **2**: 71–72.

ORVIKU K. (1960) — Nekotorye voprosy geomorfologii Estonii. Akademiya Nauk SSSR, Geomorfologicheskaya Komissiya, Moskva.

PIRRUS E. and RAUKAS A. (1996) — Late-Glacial stratigraphy in Estonia. Proc. Estonian Acad. Sc. Geol., **45** (1): 34–45.

PUURA V. and VAHER R. (1997) — Cover structure. In: Geology and Mineral Resources of Estonia (eds. A. Raukas and A. Teedumäe): 167–177. Est. Acad. Publishers, Tallinn.

RATTAS M. and KALM V. (1999) — Classification and areal distribution of glaciotectonic features in Estonia. Geol. Quart., **43** (2): 177–182.

RATTAS M. and KALM V. (2001*a*) — Glaciotectonic deformation pattern in the hummocky moraine in the distal part of Saadjärve Drumlin Field, east-central Estonia. Slovak Geol. Magaz., **7** (3): 243–246.

RATTAS M. and KALM V. (2001*b*) — Lithostratigraphy and distribution of tills in the Saadjärve Drumlin Field, east-central Estonia. Proc. Estonian Acad. Sc. Geol., **50** (1): 24–42.

RATTAS M. and PIOTROWSKI J. A. (2003) — Influence of bedrock permeability and till grain size on the formation of the Saadjärve drumlin field, Estonia, under an east-Baltic Weichselian ice stream. Boreas, **32** (1): 167–177.

RAUKAS A. and GAIGALAS A. (1993) — Pleistocene glacial deposits along the eastern periphery of the Scandinavian ice sheets — an overview. Boreas, **22** (3): 214–222.

RAUKAS A. and KAJAK K. (1995) — Quaternary stratigraphy in Estonia. Proc. Estonian Acad. Sc. Geol., **44** (3): 149–162.

RAUKAS A. and KARUKÄPP R. (1994) — Stagnant ice features in the eastern Baltic. Zeit. Geomorphologie N. F., **95**: 119–125.

RAUKAS A., KALM V., KARUKÄPP R. and RATTAS M. (2004) in press — Pleistocene Glaciations in Estonia. In: Quaternary Glaciations — Extent and Chronology, Part I: Europe (eds. J. Ehlers and P. Gibbard). Elsevier, Amsterdam.

RAUKAS A., RÄHNI E. and MIIDEL A. (1971) — Marginal glacial formations in North Estonia (in Russian with English summary). Valgus. Tallinn.

RÕUK A.-M. and RAUKAS A. (1989) — Drumlins of Estonia. Sedim. Geol., **62**: 371–384.

SMALLEY I. J. and PIOTROWSKI J. A. (1987) — Critical strength/stress ratios at the ice-bed interface in the drumlin forming process: from “dilatancy” to “cross-over”. In: Drumlin Symposium (eds. J. Menzies and J. Rose): 81–86. Balkema, Rotterdam.

SMALLEY I. J. and UNWIN D. J. (1968) — The formation and shape of drumlins and their distribution and orientation in drumlin fields. J. Glaciol., **31**: 337–390.

VAHER R. and MARDLA A. (1969) — Opyt izucheniya tektonicheskogo stroeniya uchastka Sinimäe (Severo-Vostochnaya Estonia) metodom elektrorazvedki. In: Voprosy regionalnoi geologii Pribaltiki i Belorussii (ed. F. K. Volkolakov): 119–125. Zinante, Riga.