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Applied Quaternary research in the central part of glaciated terrain

Proceedings of the INQUA Peribaltic Group Field Symposium 2006
Oulanka biological research station, Finland, September 11.–15.

Edited by Peter Johansson and Pertti Sarala

GTK

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The INQUA Peribaltic Group Field Symposium was held in September 11.–15.2006 in the central and northern part of Finland. Topics related to Late Pleistocene glaciogenic deposits in the central part of the Scandinavian ice sheet were discussed during the excursions and meeting. This publication contains 21 peer-reviewed papers that are based on the oral and poster presentations given at the symposium. The papers represent a wide range of recent Quaternary research achievements in the glaciated areas of northern Europe, America and Antarctica.

The first paper by Sarala gives a general presentation of the glacial stratigraphy and characteristics of southern Lapland and Koillismaa in Finland. In the next papers Yevzerov et al. and Korsakova & Kolka present results relating to deglaciation history in the Kola Peninsula. Depositional conditions and deposits of the last deglaciation are also discussed in the papers by Karmazienė et al. and Baltrūnas et al. for the northern and southern parts of Lithuania respectively.

In the next papers the subglacial hydrography and drainage of the melting phase are discussed. The paper by Johansson deals with meltwater action and deposits in northern Finland. In the papers by Sutinen et al. and Rattas the development of subglacial drainage systems and the related deposits are presented. Subglacial tunnel valleys and their sedimentation are introduced by Kehew & Kozlowski.

In the paper by Hang et al. varved clay studies in proglacial sedimentary environment in western Estonia are presented. Studies of buried organic sediments in the glaciolacustrine/-marine environment in Estonia are discussed by Saarse et al. In the second paper by Sutinen et al. macrofossils relating to the birch-pine-spruce succession in western Finnish Lapland have been studied. Sharapova & Semenova and Bolikhovskaya & Molodkov have done pollen investigations for palaeoenvironmental reconstruction in Kola Peninsula and Estonia. In the paper by Gyllencreutz et al. a new GIS-based reconstruction and dating database are presented.

Seismotectonic events in Lapland are discussed in the third paper of Sutinen et al. and in Kola region in the paper of Yevzerov & Nikolaeva. An example of the applied use of geophysical methods in the study of lithology and hydraulic properties of tills is given in the fourth paper of Sutinen et al. In the last papers, the urban geology of Tartu in Estonia is discussed by Kalm, and in southern Finland by Ojala and Palmu and morphotectonic investigations of the Antarctic by Lastochkin & Krotova-Putintseva.

Key words (GeoRef Thesaurus, AGI): glacial geology, deglaciation, meltwater channel, glacial features, sediments, paleobotany, Quaternary, symposia.

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Front cover: The excursion participants at the Hitura open pit.

PREFACE

The Peribaltic Group is an INQUA (International Union for Quaternary Research) research group, bringing together ice age researchers from countries around the Baltic Sea. It is one of the most active working groups of INQUA's commission meeting regularly once a year for over ten years now. The aim of the working group is to enhance the research co-operation between the countries around the Baltic Sea and to create contacts between researchers in different countries. Its most crucial concrete tasks are to compile a comprehensive file on glacial flow patterns and a GIS-based map database of late-glacial ice-dammed lakes and marginal meltwater flow systems.

The 2006 meeting of the working group took place from 11th to 15th September in Finland, having a lecture-day in 13th September at the Oulanka research station in Kuusamo, which belongs to Oulu University. All in all, 43 researchers (Fig. 1) from eight different countries came together to give lectures, discuss and share the latest research results about climate changes during the Weichselian Ice Age, the movement of glaciers, the morphological features they created and the development of the vegetation. Many lectures proposed the theory that the Weichselian Ice Age was characterised by very short glacial phases and that the phases without ice were much longer than previously thought. The new information is based on new, more accurate age determination methods and a tighter sample network. The results of recent new applications in geophysical working methods applied in Quaternary research were also introduced. The occurrence and flow patterns of glaciers of the Early and Middle Weichselian glaciation stages, however, remained open questions.

The meeting included a four-day-excursion to northern and central Ostrobothnia and southern Lapland, with the theme of 'glacial morphology in the central part of the Scandinavian ice sheet – stratigraphic and postglacial development'. Besides the key stratigraphical points in Ostrobothnia and southwest Lapland, the participants also visited Koillismaa and southern Lapland, where the glacial dynamics and the circumstances of the glacigenic deposits in the last deglaciation stage are visible. The Geological Survey of Finland and the University of Oulu's Institute of Geosciences had organised the meeting and excursion. The Finnish National Committee for Quaternary Research, the University of Helsinki's Department of Geology, the Peribaltic Group and the University of Oulu's Thule Institute also participated in organising the event.

At the meeting, 48 lectures were given, 17 of

Fig. 1. A group shot of the participants at the 2006 INQUA Peribaltic Group Meeting at the Hirvas pothole in Rovaniemi.

which were speeches and 31 poster presentations. This publication contains the written and revised abstracts of 21 lectures. The abstracts were reviewed by Professors Matti Eronen, Juha-Pekka Lunkka, Matti Räsänen and Veli-Pekka Salonen, and Doctors Heikki Hirvas and Antti Ojala. Their contribution is gratefully acknowledged. Doctors Peter Johansson and Pertti Sarala served as the reviewers and editors of this publication. Mrs Carrie Turunen kindly revised the language of many articles in this publication. Finally, the editors express their special gratitude to the Geological Survey of Finland (GTK) for supporting the publication of the workshop papers in this GTK Special Paper, volume 46. Abstracts of the lectures and excursion guide have also been published on GTK's website: http://info.gsf.fi/info/julkaisuluettelo.html under: Erikoisjulkaisut – Miscellaneous publications, numbers 55 and 56.

Peter Johansson and Pertti Sarala,
Rovaniemi 10.8.2007

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GLACIAL MORPHOLOGY AND ICE LOBATION IN SOUTHERN FINNISH LAPLAND

by
Pertti Sarala[1]

Sarala, P. 2007. Glacial morphology and ice lobation in southern Finnish Lapland. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 9–18, 7 figures.

The development of active-ice landforms from drumlins and flutings to ribbed moraines in the Kuusamo area ice-lobe indicates changing basal conditions of the glacier during the Late Weichselian deglaciation in southern Finnish Lapland. The large Kuusamo drumlin field in the eastern part of the Kuusamo ice-lobe reflects warm-based glacial conditions at the marginal part of the glacier while at the same time the central part of the lobe remained cold-based. Ribbed moraine morphology is common in the central and western parts of the Kuusamo ice-lobe.

The formation of ribbed moraines is closely related to the retreating boundary of cold- and warm-based glacier, in an internal part of the glacier margin during deglaciation.

The development of glacial landforms indicates fast flowing or even a surging type ice stream during deglaciation. A rapid climate warming at the end of the Younger Dryas has increased marginal and surface melting in the marginal parts of the ice-sheet. Simultaneously, the increase of snow accumulation in the central part of the ice lobe has caused an imbalance between the mass of the glacier's centre and the margin, which led to a fast movement of ice towards the margin.

Key words (GeoRef Thesaurus, AGI): glacial geology, glacial features, drumlins, ribbed moraines, deglaciation, glacial lobes, subglacial environment, Weichselian, Kuusamo, Ranua, Finland.

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INTRODUCTION

Glacial landforms have been successfully used to reconstruct glacier dynamics and lobe formation during the last decades. An interpretation of glacial morphology based on aerial photos and satellite images, of which the latter have been in use since the 1980's, from the entire glaciated terrain of the Northern Hemisphere. Active-ice landforms like drumlins and flutings are mainly associated with the warm-based subglacial conditions and indicate deposition in the marginal parts of actively moving ice-lobes (e.g. Aario 1990). The reconstruction of glacial processes and glacial dynamics has also been done in the central part of continental glaciers in Scandinavia and North America (e.g. Clark 1997, Punkari 1997, Kleman et al. 1997, Hättestrand 1997, Kleman & Hättestrand 1999, Sarala 2005a). The studies were mainly based on the use of subglacial transversal moraine formations like ribbed moraines and Veiki moraines with streamlined features (e.g. Hättestrand 1997).

Southern Finnish Lapland has been repeatedly situated in the central region of the last Scandinavian glaciation. Investigations for Quaternary geomorphology, sedimentology and stratigraphy have been done in many phases from the 1960's to present day in the area (e.g. Korpela 1969, Aario 1977, Kurimo 1977, Aario & Forsström 1979, Sutinen 1984, 1992, Aario 1990, Aario et al. 1997, Sarala et al. 1998, Sarala 2005a, b and c, Sarala 2006, Sarala et al. 2006). Studies were based on aerial photo interpretation, digital elevation models and test pit surveys together with analysis of spatial indicators of ice movement directions, including striae, till fabrics and the orientation of morainic landforms.

The glacial morphology in southern Finnish Lapland is dominantly composed of an assemblage of active-ice morainic landforms including streamlined features like drumlins and flutings and transverse ribbed moraines (Fig. 1) (Sarala 2005a). Large areas are also covered with relatively thick basal till deposits or thin, gently undulating till cover with a clear indication of bedrock topography (for example in the area from Ranua to the west). Based on the till stratigraphy, only two glaciation phases have existed during the Weichselian age (Sarala 2005b).

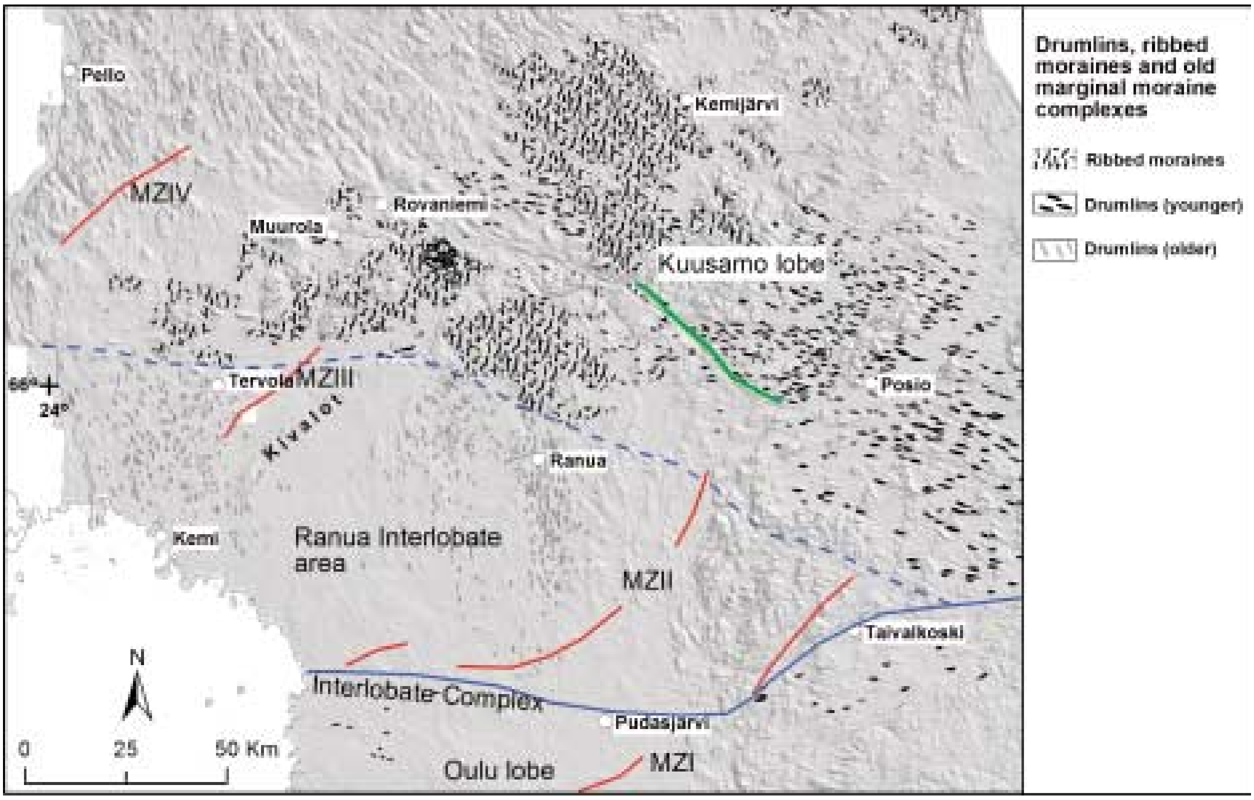

Fig. 1. The occurrence of streamlined drumlins and flutings with transverse ribbed moraines and old marginal zones MZI-MZIV (red) in southern Finnish Lapland. Digital elevation model © National Land Survey of Finland, permit No. 26/LA/07

DRUMLINS IN KUUSAMO AND RANUA

In Koillismaa area, around Kuusamo, thousands of drumlins and flutings occur forming the Kuusamo drumlin field (Glückert 1973). This drumlin field is one of the largest fields in Finland with an extent over 10,000 km^2. Ice flow has been from the northwest to the southeast during the formation of the main field. There also exists an older drumlin field below the younger one with a west-east orientation indicating an early ice flow phase (Tuoppajärvi) of the last deglaciation (Aario & Forsström, 1979).

The north-northwest to the southeast-oriented large hills and topographic depressions, presently seen as lakes and hills, are in places (for example on northern side of Taivalkoski), an indication of glacial advance before the Late Weichselian glaciation in the Kuusamo area. In the area of Ranua and towards the west, the same ice-flow phase is seen as a drumlin field. Drumlins are large with a clear indication of ice-flow from the north/northwest. There are only minor indications (western fabrics in the uppermost till and reshaped surface of some drumlins) of a later glacial advance, which overrode an older glacial morphology (Sarala 2005a, Sarala & Rossi 2006).

Glacial morphology at Oivanki (Fig. 2), northwest from the centre of Kuusamo, is composed of large, well-developed drumlins deposited during the WNW oriented (ca. 290°) ice-flow (Sarala et al. 2006). Drumlins in this area are usually long (1–2 km) and narrow (200–500 m), and both the proximal and distal ends are tapered from the ground. The distal tail is usually somewhat narrower than the proximal end. Sometimes complex drumlins, which have grown together in the middle or at the proximal end, are also seen.

A drumlin in Korkea-aho is composed of two till units (Fig. 3). The upper one is brownish-grey till with a sand matrix having a compact and mas-

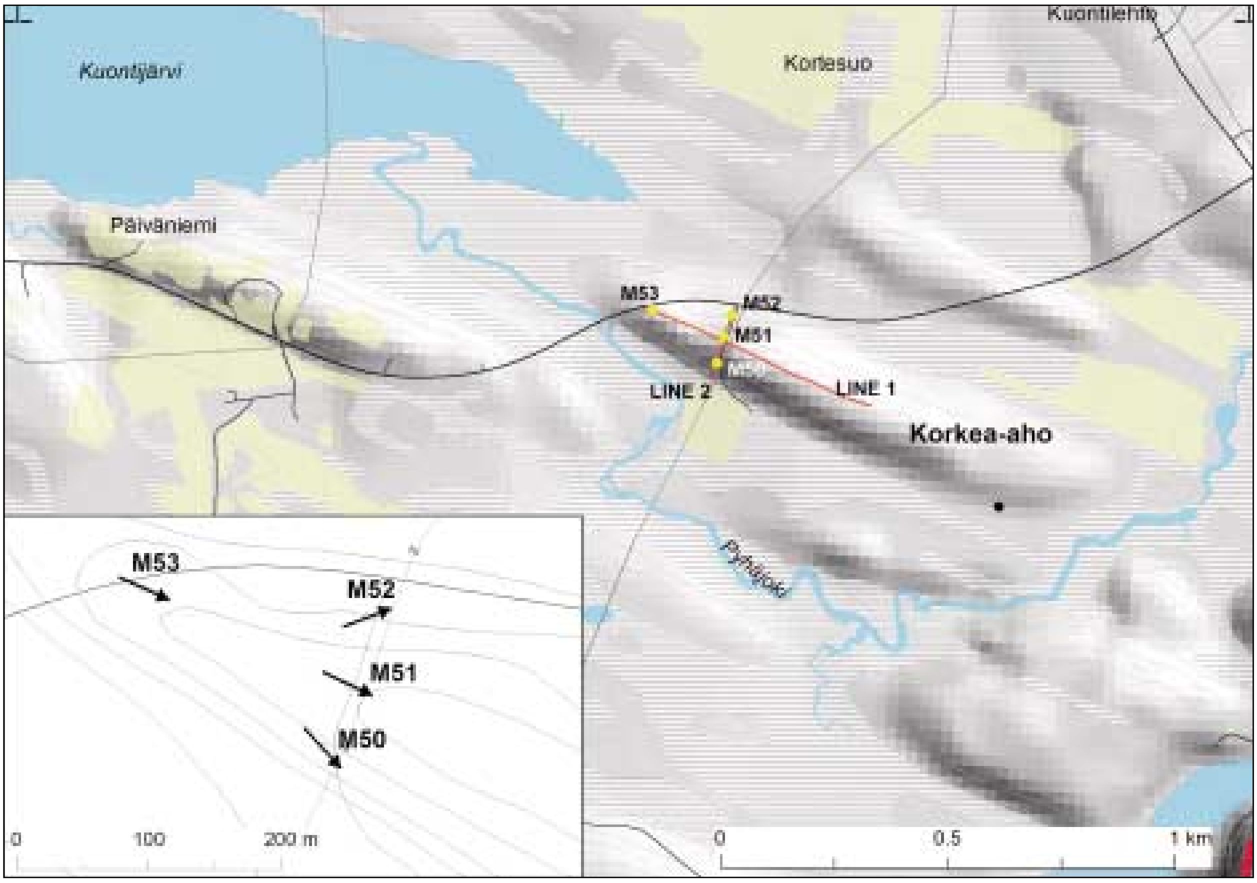

Fig. 2. Topographic map of the drumlins in Korkea-aho, Oivanki, Kuusamo. GPR (80 MHz) sounding lines 1 and 2 are marked as red lines. Till fabric of the upper till unit in different test pits (M50-M53) are presented in the lower left corner. Digital elevation model, topographic features and roads © National Land Survey of Finland, permission number 26/LA/07.

sive structure. Some fissility and thin sandy or silty stripes can be seen as a mark of subglacial deposition. The boulder and pebble content is moderate and the roundness is good. The petrographic composition of pebbles indicates long glacial transportation. Clast fabric measurements were only from this till unit in the topmost part of the ridge, showing the same orientation with a drumlin long axe (Fig. 2). Clast fabric measurements carried out in southern and northern marginal parts of the ridge showed that the stones have turned gently away from the drumlin's centre line. This is either the result of primary deposition or could also reflect secondary flow or mass-movement in the marginal parts of the ridge.

The lower till unit is composed of grey, silty or sandy till having a mostly massive structure and in places, slightly laminated parts are also seen. In a GPR (80 MHz) profile, it was possible to see that the lower till unit (Fig. 4) occurs on the southern part of the drumlin body. The older ice-flow direction has been from the northwest and till has probably been deposited during that flow phase on the stoss side of the bedrock top in the middle of the ridge. In between the tills, there is a unit of sandy/silty layers.

Fig. 3. Two till units in test pit M53 on the proximal part of the drumlin at Korkea-aho, Kuusamo. The upper till unit is composed of homogenous sandy till with fissility, typical of subglacial lodgement till. Lower till includes sorted sandy lenses and bands indicating deposition as subglacial melt-out till. Photo P. Sarala.

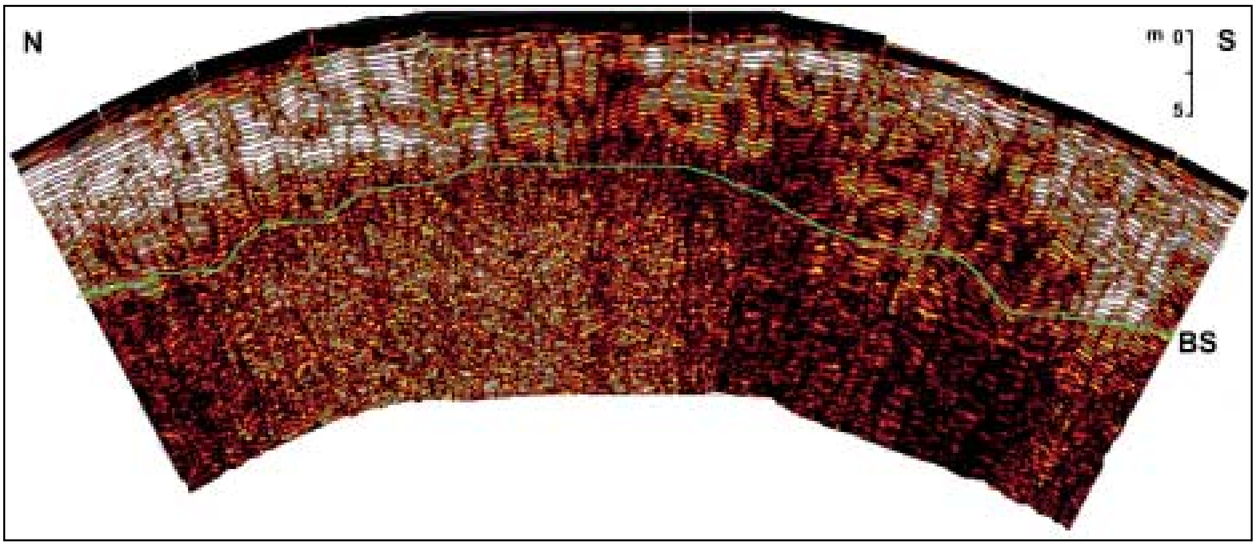

Fig. 4. GPR (80 MHz) sounding line 2 (see Fig. 2). Brownish red area above the bedrock surface (BS), on the southern side of the bedrock top is depicting lower till unit. The upper till is seen as a light grey color on the upper part of the ridge body. Between the tills occurs a unit of sandy and fines layers, sometimes also a boulder pavement.

RIBBED MORAINES IN SOUTHERN FINNISH LAPLAND

Ribbed moraines exist as uniform fields on lowland areas at the western parts of Kusamo ice-lobe and are mainly composed of Rogen moraine or hummocky ribbed moraine types (cf. Hättestrand 1997, Sarala 2003). Furthermore, a small area of minor ribbed moraines occurs in the Sihtuuna area (cf. Aario et al. 1997).

The Ranua ribbed moraine field is one of the three ribbed moraine areas in southern Finnish Lapland (cf. Sarala 2006). It was initially described by Aario (1977) and is based on his large field study data from the area. The field is composed of transverse moraine ridges, which are usually 5–15 m high, 100-150 m wide and up to one kilometre long (Fig. 5). The form of ridges is often crescentic where the edges are pointing the down-ice direction. These forms represent Rogen moraine types described by Lundqvist (1969). The hummocky ribbed moraine type (cf. Hättestrand 1997, Sarala 2003, 2006) is also common, although not with a clear indication of the ice-flow direction. Characteristically, mires and little lakes occupy depressions between the ridges (Fig. 5) and the ridge surface is mostly covered with boulders.

The ridges are composed more or less regularly of two till facies. A more densely packed lodgement and melt-out till with a fine-grained matrix occurs at the base (Fig. 6) while homogeneous melt-out and flow till occurs at the surficial parts. Pebble content increases while the roundness of pebbles decreases towards the top. The transport distance of stones is also longest at the bottom. On the surface and in the uppermost till, boulders indicate local variation of the underlying bedrock composition with very short, usually only some hundreds of meters long, transport distance. This feature has been used to trace mineralized boulders in the area (Aario et al. 1985, Aario 1990, Aario & Peuraniemi 1992). Sandy layers or stone pavements sometimes exist at the boundary between two units, although usually the contact is difficult to distinguish.

The third, bluish grey till unit (Fig. 6) is also found in places on topographic depressions, like in-between the bedrock tops. The till is clay-rich, matrix supported, compact and homogenous, indicating basal deposition, mainly by a lodgement process. Since the surface of this till is usually 6-8 m from the surface, it is hard to reach by tractor excavations but can be recognised from the ground penetrating sounding profiles (cf. Aario 1990).

Fig. 5. Part of the ribbed moraine field near the Portimojärvi village in Ranua. Test pits dug during 2005 are marked as M54-M58. Digital elevation model, topographic features and roads © National Land Survey of Finland, permission number 26/LA/07.

Fig. 6. The two lowest till units in test pit M55 at Portimojärvi, Ranua. A sharp contact between the bluish grey (bottommost) and grey tills are seen at the base of the section at a depth of about seven meters. Iron-stained, brown bands indicate the groundwater level variation. Photo P. Sarala.

GLACIAL DYNAMICS DURING THE FORMATION OF GLACIAL MORPHOLOGY

Drumlins appear to be partly erosional features but also depositional forms. The stratigraphy presented by Aario and Forsström (1979) from the Soivio area, southern Kuusamo, proved that till(s) in deeper parts of the drumlin body was deposited during the earlier Tuoppajärvi ice-flow phase, which was a single comprehensive flow unit in the Koillismaa area during the early deglaciation. The radar data from the Oivanki area (Fig. 4) shows indications of tills deposited most likely during the older glacial advance from the north-northwest. The upper parts of drumlin bodies, instead, are clearly formed by growth beneath the ice that flowed in the same direction as the lineation of drumlins. The deposition as basal lodgement and/or melt-out tills with stratified inter-till layers suggests that the subglacial conditions were mainly warm-based. The bedrock core in the case of Korkea-aho has been a shelter against the glacial erosion, thus explaining why the older deposits on the southern side of bedrock tops have been preserved.

Sarala (2006) indicated that ribbed moraine formation was a result of subglacial fracturing, mass movement and subsequent quarrying. The formation of ribbed formations initially occurred at the boundary between the frozen core and thawed or partially thawed, subglacial conditions (Fig. 7). This boundary existed about 100–150 km from the ice margin and the boundary retreated against the glacier core simultaneously with the marginal melting. Subsequently, after the fracturing and movement of subglacial ice-sediment blocks variable pressure conditions under the moving ice-sheet initiated the freeze-thaw process that caused the subglacial quarrying and deposition of local bedrock material on the ridge tops.

The glacial morphology in southern Finnish Lapland is indicative of ice-lobe development during the latest melting phase of the Scandinavian Ice-Sheet. As deglaciation continued and the ice margin reached the area of the modern border in eastern Finland, rearrangement occurred and the ice-lobes of Kuusamo and Oulu were formed. The Oulu lobe in the southern parts of the area contin-

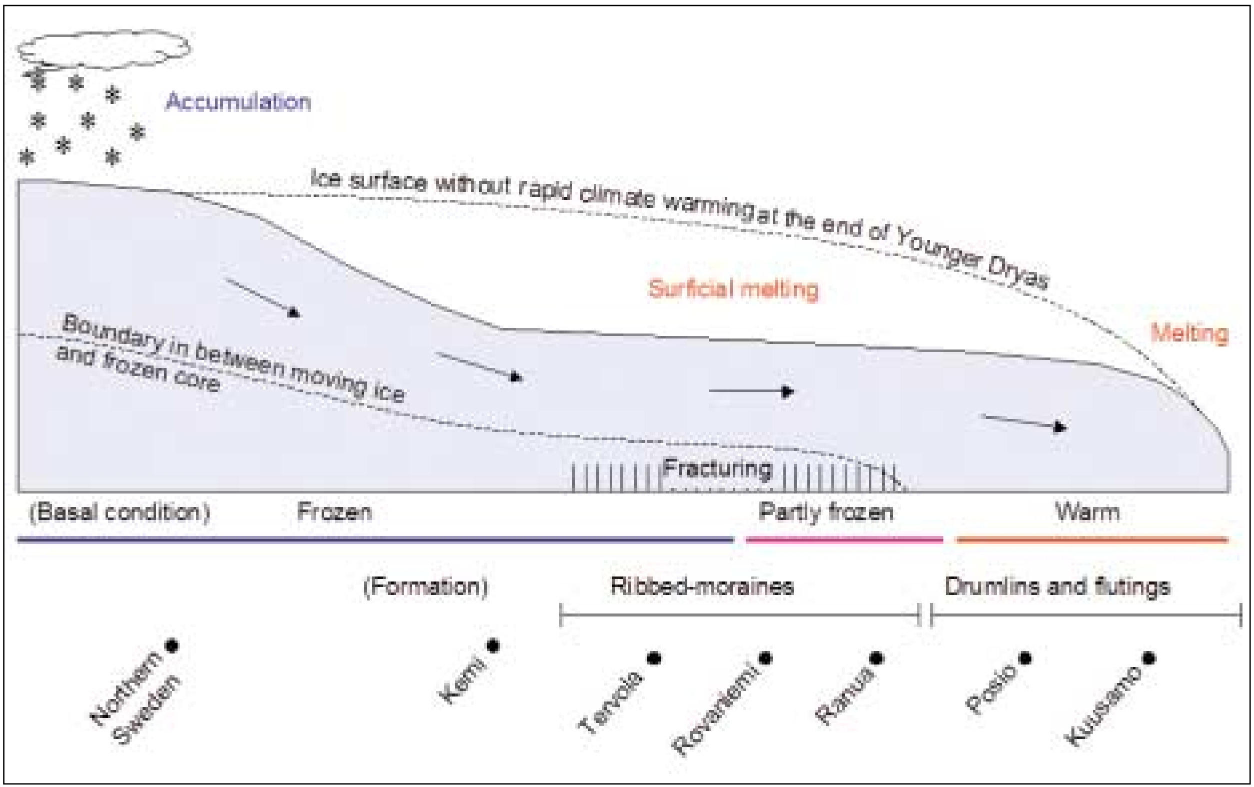

Fig. 7. A rapid marginal melting and precipitation with a subsequent increase of accumulation in the central parts of the glacier has sped up the flow rate of the Kuusamo ice-lobe. The ribbed moraine formation seems to be closely related to the retreating boundary of the cold- and warm-based glacier.

ued flowing from west to east but in the north, the flow direction of the Kuusamo lobe was turned from the northwest towards the Oulu lobe. A large Pudasjärvi-Taivalkoski-Hossa Interlobate Complex was initially formed into the contact between the Kuusamo and Oulu lobes (Aario 1977, 1990, Aario & Forsström 1979) but later on, at the second phase, at the contact between the Oulu lobe and the southern margin of Ranua Interlobate area (Sarala 2005a).

A large Kuusamo drumlin field in the east indicates warm-based glacial conditions at the outer margin of Kuusamo ice-lobe. In the central and western parts of the ice lobe area, the ribbed moraine fields of Ranua, Tervola and Kemijärvi were formed. The ribbed moraine formation was interpreted to be closely related to the retreating boundary of cold- and warm-based glacier, in an internal part of the glacier margin during deglaciation (Kleman & Hättestrand 1999, Sarala 2005a). A rapid marginal melting, increased precipitation and accumulation in the central parts of the glacier increased the flow rate of the Kuusamo ice-lobe (Fig. 7). Thus, the development of active-ice landforms from the eastern part of the Kuusamo ice-lobe to the west is indicative of fast flowing, even a surging-type ice stream during deglaciation (cf. Hart 1999).

AGE OF DEPOSITION

Based on the latest dating results, the build-up of glacial morphology in southern Finnish Lapland can, in most cases, be dated back to the Middle and Late Weichselian (Mäkinen 1999, 2005 and Sarala 2005a and b). The Middle Weichselian glacial advance was the first to cover the southern Finnish Lapland area after the Eemian interglacial. The Early Weichselian glacial advance(s) was probably so weak that it affected only the area of central and/or northernmost Finnish Lapland (cf. Svendsen et al. 1999, 2004, Siegert et al. 2001, Arnold & Sharp 2002).

The rapid warming of the climate at the end of the Younger Dryas (12.8-11.6 ka ago) (cf. Boulton et al. 2001, Bard 2002) increased mean temperature from 8–10°C during a few hundreds of years, causing a strong melting of the glacier margin and surface (Fig. 7). An imbalance between the warm surface and cold-based part of the moving ice sheet followed and might be the reason for the initiation of ribbed moraine formations in the core areas of the latest glaciers in the northern hemisphere (Sarala 2005a). Similar rapid climate change has not happened since, so it could be an explanation why the ribbed moraines are not present in modern glaciated areas.

The Ranua interlobate area in between the Kuusamo and Oulu lobes with the north-northwest to south-southeast oriented drumlin field is a relict from the older, probably the Middle Weichselian glacial phase (Sarala 2005a, Mäkinen 2006) and has been preserved under the cold-based, central area of the Late Weichselian glaciation.

DISCUSSION AND CONCLUSIONS

This study was based on the interpretation of glacial landforms with stratigraphical studies in the test areas in Kuusamo and Ranua, southern Finnish Lapland. As several earlier studies prove, the active ice landforms are useful in reconstructing glacial dynamics during deposition. Landforms like drumlins, flutings and ribbed moraines indicate the presence of active ice-lobes after the separation of the ice-sheet margin during deglaciation. In many cases, the interpretation of glacial morphology is difficult due to the existence of interlobate areas between the active lobes. As a relict of the cold-based glacier core and having only minor marks of basal movement, older glacial morphology has been preserved under passive interlobate areas (cf. Sarala 2005a). Thus, the glaciogenic landforms of different ages can be found in the same area or even overlapping each other.

The glacial morphology in southern Finnish Lapland is dominantly composed of assemblages of active-ice morainic landforms including streamlined drumlins, flutings and transverse ribbed moraines. These forms are the dominant landform types in the area of the Kuusamo ice-lobe and indicate the Late Weichselian glacial morphology. Instead, the Ranua interlobate area with the north-northwest to south-southeast oriented drumlin field is a relict from the older, probably Middle Weichselian glacial phase and was preserved under the cold-based, central area of the Late Weichselian glaciation. According the latest dating results (Mäkinen 2005, Sarala 2005a and b), the Middle Weichselian glacial phase was the first one to cover the southern Finnish Lapland area after the Eemian interglacial.

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DEGLACIATION OF THE RYBACHI AND SREDNI PENINSULAS IN THE LATE PLEISTOCENE, NORTHWEST RUSSIA

by
Vladimir Yevzerov[1], Jakob Møller[2], Vasili Kolka[1] and Geoffrey Corner[3]

Yevzerov, V., Møller, J., Kolka, V. & **Corner, G. 2007**. Deglaciation of the Rybachi and Sredni Peninsulas in the Late Pleistocene, northwest Russia. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 19–24, 3 figures, 1 table.

The position of the high marine limit was determined on the basis of the highest altitude of marine deposits and relief forms related to the sea activity. In most locations, these were adjacent beach ridges and outwash plains located on the distal side. In addition to the authors' original measurements, data from Tanner (1930) were also used. A detailed study of the position of locations of the marine limit and marginal moraine ridges has made it possible not only to reconstruct the history of gradual glacial shrinking and the retreat of the ice margin, but also to identify the periods of invert motion of the glacier, which occurred during the Late Pleistocene due to coolings.

Key words (GeoRef Thesaurus, AGI): glacial geology, deglaciation, sea-level changes, highest shoreline, shore features, end moraines, Pleistocene, Kola Peninsula, Russian Federation.

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INTRODUCTION

The Kola region includes the Kola peninsula and adjacent shelves of the Barents and White seas. Two deglaciation stages have been distinguished in this area (Yevzerov 1998). In the first stage, which lasted approximately 5,000 years (Yevzerov & Nikolaeva 2000), aerial deglaciation took place. Three marginal belts were formed during the period between the Last Glacial Maximum in the Kola region (near 16,000 years BP) and the Preboreal. During interstadial warming (before the Oldest Dryas, Bølling and Allerød), extensive peripheral covers were cut off from the main ice mass and thick glaciofluvial deposits accumulated only in the periglacial basins near the edge of the active ice. The glacier that advanced during the phases of stadial coolings (Oldest Dryas, Older Dryas and Younger Dryas) deformed interstadial deposits and built mainly push moraine ridges. In the second stage of deglaciation, the rest of the ice sheet was dissected by extended marine gulfs during the warming in the beginning of the Holocene. In these gulfs, glaciomarine sediments were superseded by marine deposits, probably in connection with the final melting of the ice about 9,000 years BP.

Two bands of the push moraine ridges are found in the northwest portion of the Kola region. The southern band is a continuation of the Tromsø -Lyngen regional formations of Norway (Marthinussen 1974). In the west, the formation crosses the Pechenga River about 5 km from the mouth and approaches the Murman coast near the Zapadnaya Litsa Bay in the east. Ridges of this band were generated during the Younger Dryas and are part of an inner band of marginal belt II. This band was generated in the Kola region during the Bølling (Yevzerov & Nikolaeva 2000). The second band is an advanced fragment on the Sredni Peninsula and includes push moraine ridges in Ejno Bay and in the northern branch of Motovsky Bay (Bolshaya Motka Bay). These ridges were generated in connection with a cooling during the Older Dryas. Between these bands, a fragment of the dump moraine ridges band is found about 2.5 kms from the mouth of the Pechenga River. It could be formed only during the Bølling. Absence of an extended band of dump moraine ridges indicates adverse conditions for the occurrence of periglacial basins in the studied area. The surface of the ice sheet bed was inclined in the direction of the ice movement and was cut by rivers valleys flowing into Barents Sea. The shelf of the southeastern part of the Barents Sea was exempted from glacial ice during the Bølling (Polyak & Mikhailov 1996). The shelf near the Rybachi and Sredni Peninsulas was certainly free from glacial ice during the Bølling and a dead ice cover bordered the main ice mass of active ice on the peninsulas at the same time.

STUDY AREA

Our research is a continuation of studies of coastal and glacial deposits of Rybachi and Sredni Peninsulas that was started by Tanner (Tanner 1930). These peninsulas are situated in northwest Russia, near the boundary between Russia and Norway (Fig. 1) and are a low-relief undulating plateau dissected by numerous creek and river valleys. Most of the area is located at an altitude of about 200 m above sea level (a.s.l.). The eastern part of Sredni, however, is more elevated, with the highest altitudes exceeding 300 m a.s.l. The surface of the peninsulas is covered by a thin, discontinuous layer of diamicton consisting of decomposition products of Riphean rocks: sandstone, siltstone, clay shale, dolomite and conglomerate. Outcrops of these rocks are common. The moraine is scarce, occurring largely in the southern part of the study area. Erratic boulders of gneiss, granite and basic rocks of different ages (Archaean and probable Proterozoic), however, are abundant. The boulders were moved more than 30 km from the southwest to the northeast based on the orientation of glacial strias. Marginal ridges of push moraines have been found in the northern branch of Motovsky Bay (Bolshaya Motka Bay) and in the Eina Bay, on the southern coast of Rybachi. The valleys of many rivers and creeks, facing east and north, are partially covered by glaciofluvial delta deposits. On both peninsulas, Pleistocene and Holocene coastal sediments are abundant, locally reaching an altitude of about 100 m a.s.l. Some idea as to their distribution can be gained from Figure 1.

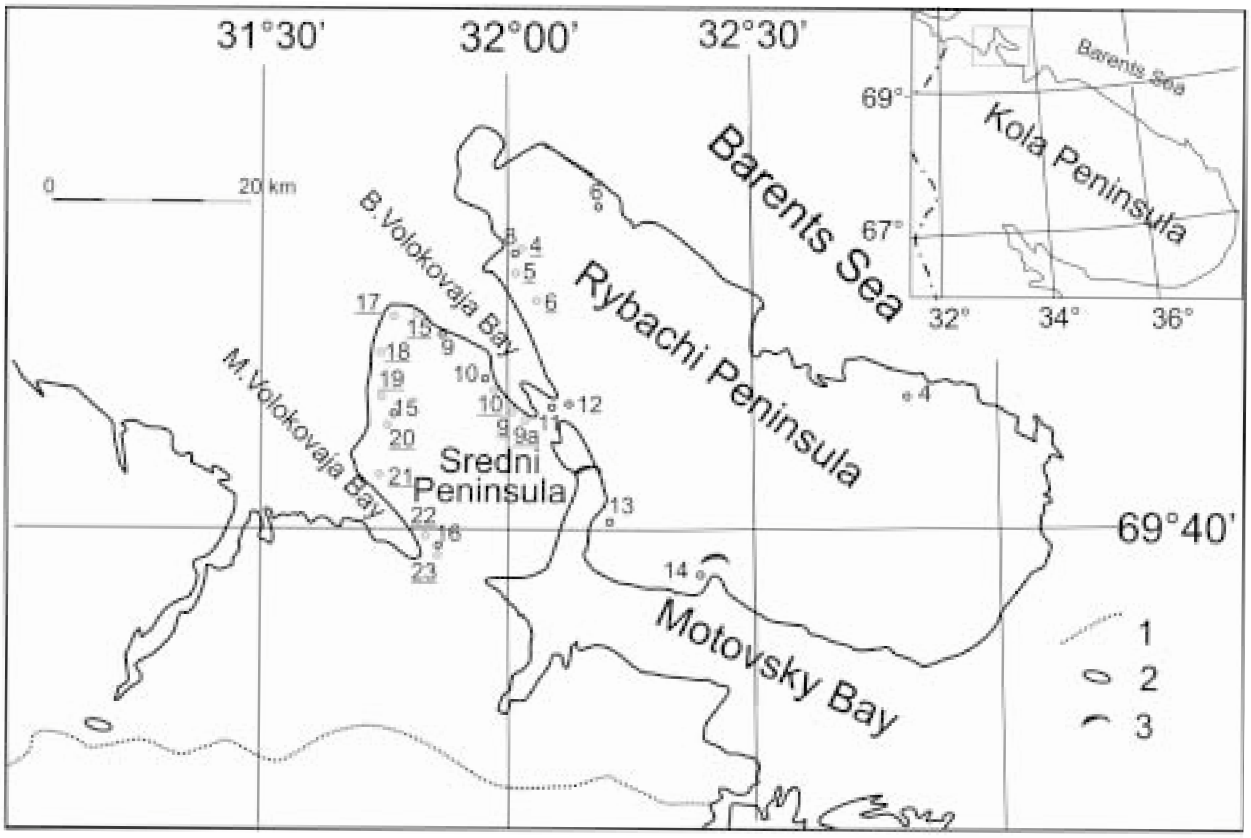

Fig. 1. Sketch-map of the study area showing the location bands of dump and push moraine ridges, the location of the levelled marine limit localities, projection plane A-A1 for the equidistant shoreline diagram perpendicular to isobases of uplift and presented in Fig. 2. Hereinafter the numbers of locality points from Tanner (1930) are underlined.

THE POSITION OF THE HIGH MARINE LIMIT

The position of the high marine limit (HML) was determined on the basis of the highest altitude of marine deposits and relief forms related to the activity of the sea. In most locations, these were adjacent beach ridges and outwash plains located on the distal side. In this chapter, we shall discuss the results from the study of the extent and altitude of the marine deposits and relief forms that are confined to the marine limit of the Late Pleistocene. The altitudes were measured with the use of a theodolite from triangulation points or from the sea level with subsequent recalculation to the average sea level. In addition to the authors' original measurements, data from Tanner (Tanner 1930) was also used. The locations studied are shown in Fig.1, and the altitudes of the marine limit are given in the Table1.

The corrections of measured beach feature altitudes approximately reflecting the energy of waves in each point were included in the measurements results. The magnitude of the corrections was defined by measuring the altitudes of crests of modern beach-ridges on the Rybachi and Sredni Peninsulas. All beach ridges in the bays occur at an altitude of more than 2 m above the average sea level, and those on the open sea coast are commonly located no higher than 6-7 m above this level. Therefore, the correction range was 2-6 m. Differentiation of the corrections within this range was performed with consideration for the coast morphology, although the precision is unlikely to exceed one meter. Consequently, the resulting altitudes of the marine limit given in Table 1 have a precision of no more than ± 1 m. The resulting altitudes of the marine limit were used to construct an equidistant shoreline diagram shown in Fig. 2. The line S0, plotted with the use of the authors' data, corresponds to the Late Dryas. As to other lines on the diagram, the greater inclination to the abscissa, the earlier geologic events they indicate. This trend, in conjunction with the spatial position of the locations that were used to construct the diagram, and the data on marginal moraine ridges, make it possible to reconstruct the succession of deglaciation in the study area.

Table 1. The altitude of the marine limit on the Rybachi and Sredni Peninsulas

Loc. No	**Locality**	**Type of feature**	**Height abo¬ve mid sea level in m**	**Correction in m**	**Adjusted height in m**
2	Tsypnavolok	Ridge	99.9	4	95.9
4	Chorhy Cape	- „ -	99.8	4	95.8
6	Skarbeevskaya Bay	- „ -	92.5	5	87.5
8	Korovy Cape	- „ -	87.2	2	85.2
4[1])	Klupuniementunturi	Abrasion. terrace	88.7	2	86.2
5	Niittymukka	Ridge (?)	101.5-102.9	5	96.5-97.9
6	Syväjoki	Abrasion. terrace	86.4	2	84.4
11	Klubb Mauntin	Ridge	94.6	5	89.6
9a	Kohmelo-kuru	Abrasion. terrace	94.49	2	92.5
12	Ozerko	Ridge	90.0	2	88.0
9	Perähamina	Ridge (?)	95.21	3	92.2
10	Ittelin	Ridge	93.9	5	88.9
10	Kuivatunturi	- „ -	94.99	5	90.0
9	Zemlyanoe	- „ -	92.6	4	88.6
13	Monument	- „ -	95.7	5	90.7
14	Eina Bay	- „ -	89.4	2	87.4
15	Vähä-Outa	Abrasion. terrace	95.93	6	89.9
17	Pummankitunturi	- „ -	96.0	5	91
18	Poropellot	Ridge (?)	104.3-107.7	6	98.3-101.7
19	Isomukka	Abrasion. terrace (?)	99.1	6	93.1
15	Matalaniemi Bay	Ridge	97.7	5	92.7
20	Govcce`rotto	(?)	99.79	6	93.8
21	Porttu`njard	Ridge	96.5	4	92.5
22	Palve`tuo-ddar	(?)	95.9	3	92.9
16	Sputnik	Ridge	90.7	3	87.7
23	Madde`muetke	- „ -	91.0	3	88.0

[1]) *The numbers of the localities from Tanner (1930) are underlined.*

THE SUCCESSION OF DEGLACIATION IN THE AREA

The history of deglaciation of Rybachi and Sredni Peninsulas includes some stages. All the stages are indicated in Fig. 3. During the first two stages, the ice sheet margin gradually retreated towards the southwest and possibly southeast, leaving the northeastern and northwestern part of Rybachi and probable Motovski Bay ice-free. The position of the locations by which the marine limit was defined and the two stages were determined can be fairly well correlated in situ and on the diagram (Figs. 1, 2 and 3). During the third stage, the glacier advanced and joined as a part of the ice sheet of the previous stage (Fig. 3:3). The occurrence of a new expansion of the glacier is supported, directly or indirectly, by the presence of push moraine ridges in the northern branch of Motovsky Bay and in Eina Bay (north of locality 14), and the above mentioned information on the history of deglaciation of the Kola Peninsula during the Late Pleistocene. This advance of the glacier probably was caused by the cooling during the Older Dryas. In the following stage (Fig. 3:4), all territory of the Rybachi Peninsula and a northwest part of the Sredni Peninsula were released from ice. A small time interval probably existed in the formation of the HML marked with points on lines 3-4 of diagram (Fig. 3). Deglaciation by position of point 14 from this community (Fig. 2) took place definitely after the formation of push moraine ridges during the Older Dryas and hence, it is connected to the subsequent warming of the Allerød. In this period, there was a strait between the Rybachi and Sredni Peninsulas during the HML on Mount Klubb (No 11 in the table). It exceeds the altitudes of the low sites of an isthmus to the east and a southeast from the mentioned mountain. Thus, the peninsula Rybachi has turned into an island. Ice was probably kept active for some time in the southwest part of Sredni Peninsula at the end of the fourth stage. However, evidence of it are absent and in relation to Fig. 3, IV dead ice is only shown. All territory of Sredni Peninsula was released from ice during the Allerød in the three following stages. Positions of dead ice massif at the end of each of these stages are shown accordingly on Fig. 3, 5, 6 and 7. A new approach of the ice sheets in the Younger Dryas has not reached the Rybachi and Sredni Peninsulas. The border of the maximum distribution of active ice in this period is seen much further to the south (Fig. 1).

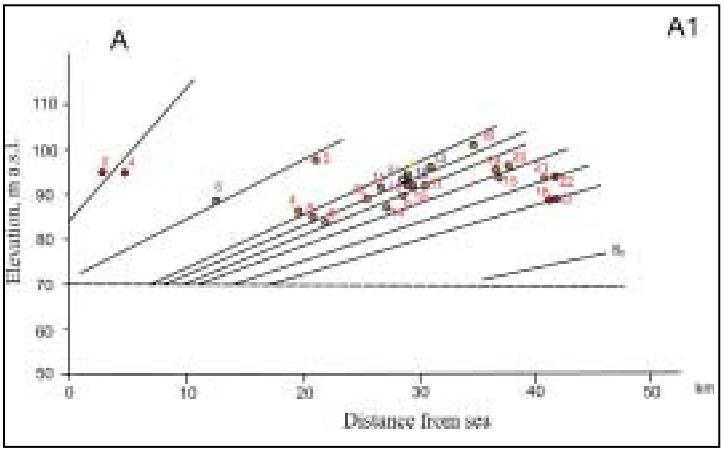

Fig. 2. Equidistant shoreline diagram for Rybachi and Sredni peninsulas constructed on the basis of the HML in different points. Heights of the marine limit shown in the Table 1. The line S0 corresponds to the Younger Dryas. It put according to authors. Other lines correspond to the surfaces that were formed by the sea before the Younger Dryas. They are inclined to a line S0 under the same corners as on the diagram of M. Marthinussen [1] for the stripe from Varangerfjord and the east of the Varanger peninsula till area of Inari lake.

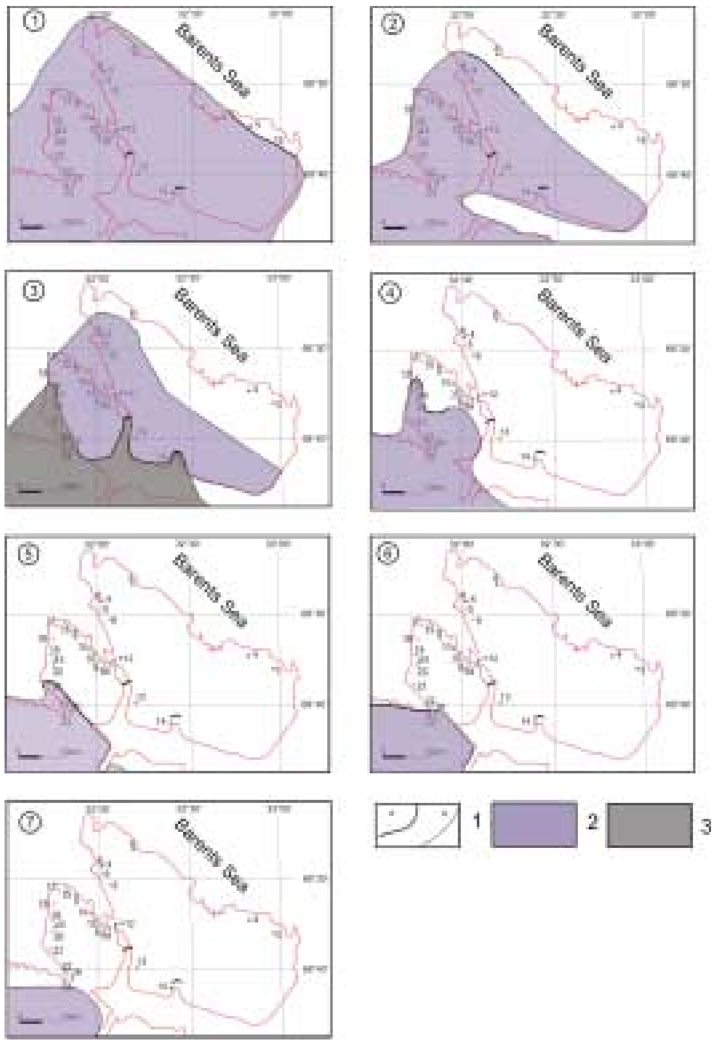

Fig. 3. The succession of the Rybachi and Sredni peninsulas deglaciation. 1- border of ice distribution established (a) and assumed (b); 2 – dead ice; 3 – active ice of ice sheet.

CONCLUSIONS

Deglaciation of the area of the Rybachi and Sredni Peninsulas started in the northeast and finished in the southwest. During deglaciation, there was an inverted motion (advance) of the glacier due to cooling. The deglaciation of the study area started during the Bølling and ended during the Allerød. The detailed study of the position of the marine limit and marginal moraine ridges has made it possible not only to reconstruct the history of gradual shrinking of the glacier and the retreat of the ice margin, but also to identify the periods of inverted motion of the glacier, which occurred during the Late Pleistocene due to cooling.

ACKNOWLEDGEMENTS

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LANDSCAPE AND GEOLOGICAL FEATURES OF TECTONIC ZONES IN THE KHIBINY MOUNTAINS, KOLA PENINSULA, NW RUSSIA

by
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Korsakova, O. & **Kolka, V. 2007.** Landscape and geological features of tectonic zones in the Khibiny Mountains, Kola Peninsula, NW Russia. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 25–30, 4 figures.

Tectonic zones impose the borders (lineaments) of the Khibiny block structures. Under uplifting conditions the tectonic zones are defined as weakened zones, where the rocks have undergone intensive exogenic processes. The geological features of the tectonic zones are the linear weathering crust and reworked preglacial deposits. The direct relationship among the spatial location of the valleys and tectonic zones illustrates that the multiordered river valleys are the main landscape feature of tectonic zones. In the interstream areas, indicators of the tectonic zones are seen as gorges, fractures, and low-amplitude ledges. Other indicators are zones of disintegrated rocks; glacial cirques and corries developed along tectonic zones; permafrost landforms; ground water discharges at the bottom of the graben-like gorges and along the ridges' slopes; the firns in cirques and fractures; and linearly located soil and vegetation on the plateau-like divides.

Key words (GeoRef Thesaurus, AGI): tectonics, block structures, lineaments, landforms, valleys, gorges, weathering, Khibiny Mountains, Kola Peninsula, Russian Federation.

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INTRODUCTION

From numerous investigations (Onokhin 1975, Mel'nikov 2002), the block structure of the Khibiny massif and its high tectonic activity have been established. Seismic data derived from earthquake investigations (Mel'nikov 2002) and instrumental measurements indicate that the present-day tectonic uplifting reaches 2-4 mm/year in this area (Mescheryakov 1973). Tectonic zones impose the borders (lineaments) of the Khibiny block structures and they inherit numerous features of different forms. Near-surface parts of the Earth's crust and land surface tectonic zones have landscape-geological indicators, manifested in weathered rocks and/or specific landscape forms.

STUDY AREA

The Khibiny Mountains (highest point at 1200 m a.s.l) are situated in the central part of the Kola region (Fig. 1). It is a concentrically zoned multiphase intrusion composed of nepheline syenites and to a lesser amount, of ultrabasic alkaline rocks. The main orographic feature of the Khibiny Mountains is their table-like shape. Deep canyons, U-shaped and wide river valleys cut through the massif. Some of the mountains are capped with extensive plateaus, and encircled with cirques and steep slopes. The depth of the large cirques reaches 400 m. Cutting valleys can reach a depth of 800 m.

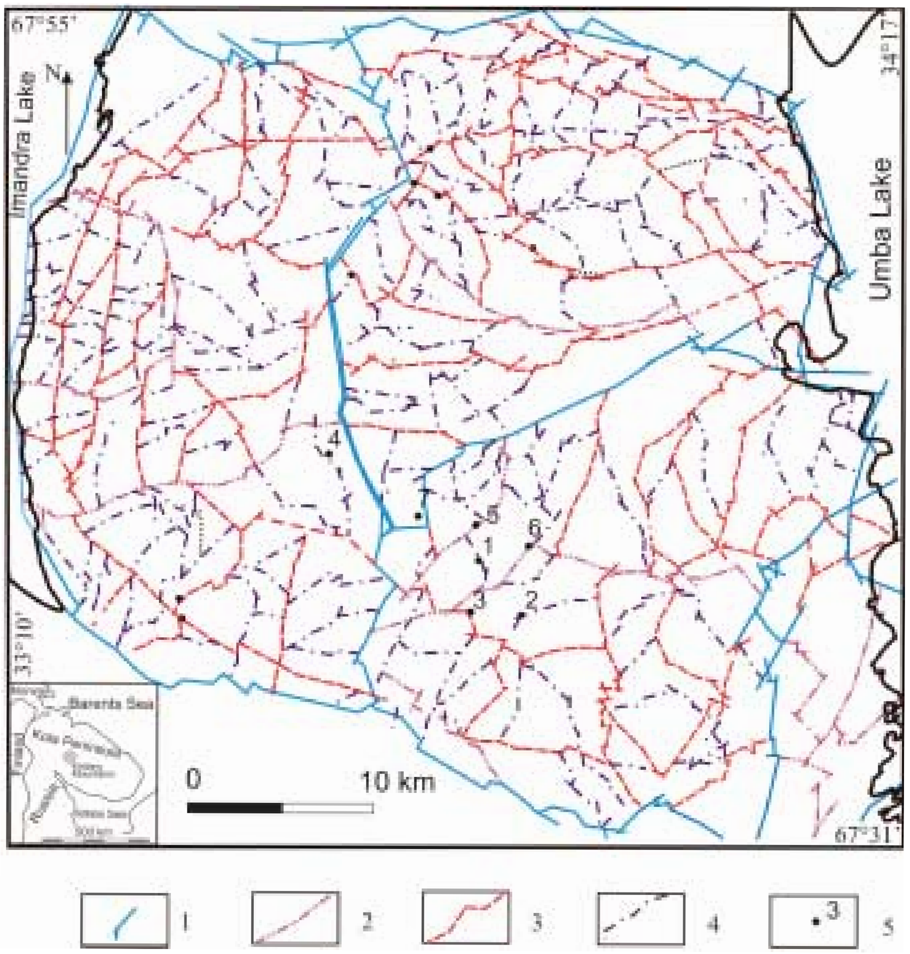

Fig. 1. Location of the study area and scheme of the lineaments of the Khibiny Mountain massif (lineaments of 5th order are not included). 1 – lineaments of the 1st order, 2 – of the 2nd, 3 -of the 3rd, 4 – of the 4th; 5 – the locations of observed weathering rocks, deposits (above-mentioned in the text are marked by numbers – saddle between Bolshoy Yuksporr and Maly Yuksporr plateaus (1); Apatity (2) and Pod'emny (3) cirques, valley of Poachvumjok (4) and Saami (5) Rivers, Juksporlak mountain pass (6), Bolshoy Vud'javr Lake (7)).

METHODS

To establish the landscape and geological features of the tectonic zones in the Khibiny Mountains, morphometric, remote sensing and fieldwork methods were applied to study the relief and composition of the rocks. Based on decoding of the aero- and satellite images, and of the morphometric analysis, a sketch map of the multi-ordered lineaments (tectonic zones) was compiled. Spatial position and order of lineaments was defined in accordance with the order of the river valleys developing along the tectonic zones. During the field work, a verification of distinguished lineaments was done and landscape features of tectonic zones beyond river valleys were defined. Published data were also used.

The main theoretical prerequisite for distinguishing the tectonic zones is based on the assumption that the lineaments are surface expressions of tectonics breaking the earth crust. In this case, the position of the lineaments is outlined mainly by the river valleys and other negative topographic forms. The river valleys are developed along the tectonic zones by selective erosion, for example, the order of the river valley depends on the intensity and depth of the tectonic zone responsible for its formation. In the water divides, the lineaments are traced by the alternate negative topographic forms and are manifested in the rock composition. The five orders of the lineaments were established within the Khibiny Mountains (Korsakova et al. 2005; Fig. 1).

RESULTS AND DISCUSSION

The Khibiny massif represents the ordered hierarchic system of structural elements, named as tectonic blocks. The block structure is conditioned by the development of the multirange discontinuities (from microfracturing and contacts between the crystals to faults, extending hundreds kilometers). These discontinuities are caused by the interaction of the main geodynamic processes: continental drift, tectonic activity, and elastic resistance of the Khibiny massif rock. These factors define the hierarchic field of natural tensions and cause the destruction of the massif due to disconnection and displacement along the structural discontinuities of different orders that segregate the Khibiny massif and form tectonic blocks of different ranges. The natural tensions and strength characteristics of the structural forms are determined by the dimension of the related volumes (i.e. by the dimension of the block structures of different orders).

The rocks of northeast Fennoscandian Shield are under horizontal crustal compression because of adjacent ocean-floor spreading and continental drift. During the Mesozoic and Cenozoic, the tectonic uplift of the Khibiny massif resulted in the partial loss of horizontal compression existing deep in the Earth's crust, leading to the extension of the rising rock masses. Due to these processes, non-uniformly scaled vertical cracks are revealed and the sheeting of thin rocks occurs in near-surface parts of the Earth's crust. In other words, the tectonic zones form. Within these tectonic zones the rocks are loosened and shattered. The higher erosion intensity occurs in the tectonic zones because disintegrated rock masses are subjected to denudation-accumulation processes.

Geological features of tectonic zones are presented by the linear weathering crust, preglacial sediments (breccia, conglomerate, gravelite, clay), weathered and disintegrated hard rock that were observed in exposures and revealed by drilling under the glacial sediments during the prospecting and exploration within the Khibiny Mountains (Armand 1964, Mel'nikov 2002). Disintegrated hard rock contains schpreustein and epigenetic minerals – hydromica, nontronite, hydrohematite, limonite, chalcedony, anatase.

Distribution of the disintegrated rocks, linear weathering crust and preglacial sediments is confined to the river valleys, gorges, hollows and saddles in interstream areas. For example, the linear zones of shattered rocks were found in the saddle between Bolshoy Yuksporr and Maly Yuksporr plateaus, in Apatity and Pod'emny cirques and in adjacent interstream areas along the deep saddles in the valley of Poachvumjok River. All these zones are traced by the lineaments of the 4th and 3rd orders (Fig. 1, points 1, 2, 3, 4, correspondingly). The 2nd order lineaments are seen in the intensively fractured and altered rock with limonitized apatite, underlying the glacial sediments and ancient thick diluvium in the valley of Saami river (Fig 1, point 5; Fig. 2), the fractured altered ryschorites revealed by drilling and trenching in Juksporlak mountain pass (Fig. 1, point 6). The melteygite fragments with gouges

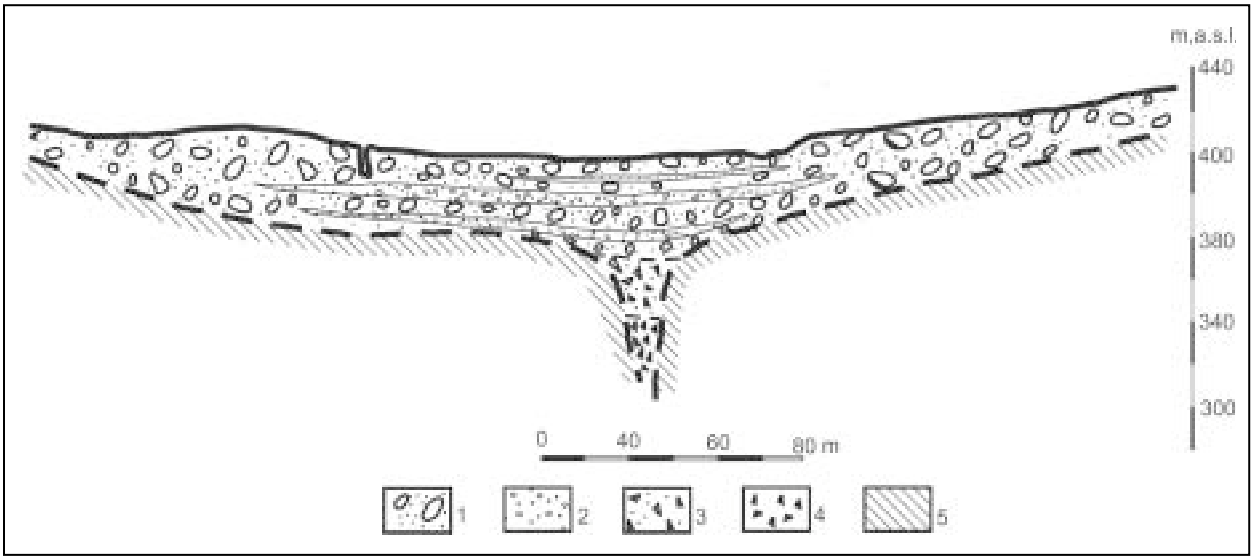

Fig. 2. Geological section on the valley of the Saami River (after Armand, 1964; location is shown on Fig. 1 as point 5). 1 - till, 2 - glaciofluvial sand (fine and middle size grains) with pebbles and boulders, 3 - slide eluvium rock, 4 - intensively fractured and altered rock with limonitized apatite, 5 - bedrock.

Fig. 3. Landscape features of tectonic zones on the interstream areas expressed in the relief and ground: a - graben-like gorges, fractures; b - low-amplitude ledges of topographic surface, c - linear zones of highly disintegrated rocks, d - permafrost land-forms (solifluction bands).

of clayey rocks were found in the drill core at the bottom of the Bolshoy Vud'javr Lake, situated along the lineament of the 1st order (Fig. 1, point 7).

The available geological data indicate that tectonic zones are expressed as linear segments of disintegrated rocks. They coincide with negative land forms, which represent lineaments (block divides or tectonic zone) of different orders.

The landscape features of tectonic zones are variably expressed on the surface. The link between the negative topographic forms and the fracture system of the alkaline rocks through the intermediate disintegrated zones and the linear weathering crust explains the radial-concentric stream pattern within the Khibiny Mountains The direct relationship between the spatial location of the valleys and tectonic zones is confirmed by model constructions (Mel'nikov 2002, Korsakova et al. 2005) allows multiordered stream valleys to be considered as their main landform feature.

The interstream areas in Khibiny Mountains are seen mainly in the plateau-lake surfaces. The plain usage of morphometric method is able to only approximately determine the lineaments position here. In this case, the indicators of tectonic zones are precisely identified during field investigations.

The most evident indicators of tectonic zones within the interstream areas are the negative structural forms and relief elements (graben-like gorges, fractures (Fig. 3a) and low-amplitude ledges of topographic surface (Fig. 3b)). As a rule, these indicators are characteristics of the block divides of the lowest (the 5th) order. The structural forms of higher order tectonic zones have a larger set of landscape features. During the field investigation, the different landscape features of tectonic zones were also established: the linear zones of highly disintegrated rocks in the plateau-like interstream areas (Fig. 3c); permafrost landforms (stone polygons, solifluction bands (Fig. 3d)) spatially confined to linear zones of highly disintegrated rocks; ground water discharges at the bottom of the graben-like gorges and along the slopes of ridges (Fig. 4a); the linearly localized soil and specific vegetation (Fig. 4b); the firns forming in ancient cirques and in the younger grabens and fractures (Fig. 4c).

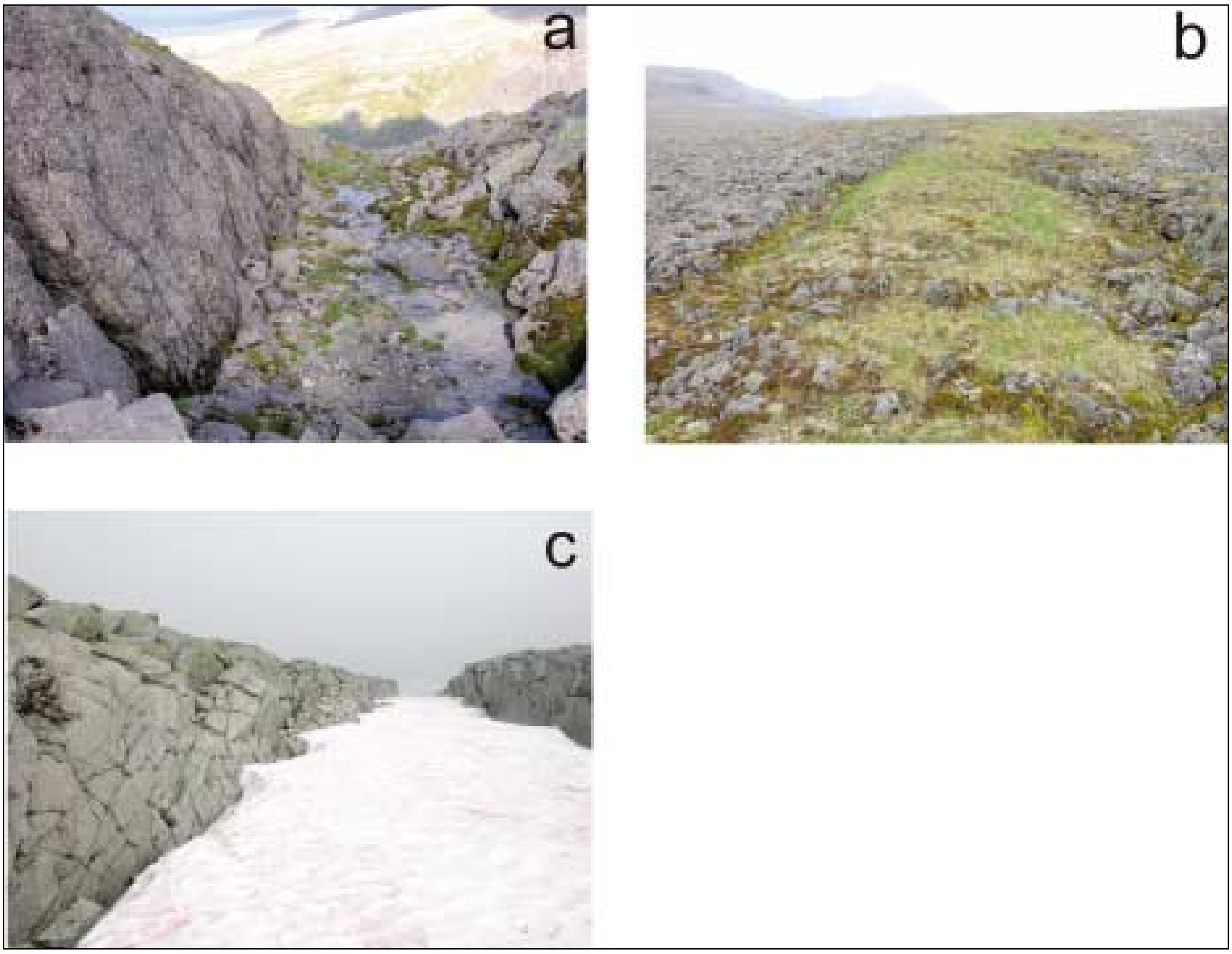

Fig. 4. Landscape features of tectonics zones on the interstream areas expressed in the peculiarities of waters, soil, vegetation: a - ground water discharges at the bottom of the graben-like gorge, b - linear localization of the soil and vegetation, c - firns in the younger graben.

CONCLUSIONS

Subjected to the prevailing uplifting conditions, the tectonic zones are defined as weakened zones where the rocks are intensively affected by exogenic processes. Location of the zones defines the development of the majority of negative landforms in the Khibiny Mountains. The main geological features of the tectonic zones are the linear weathering crust, altered preglacial sediments confined to negative landforms. The manner and localization of exogenic processes (erosion, nivation, slope outwash, permafrost solifluction and others) are also defined by the location of tectonic zones. The tectonic zones have geological, geomorphological, hydrogeological, edaphic and geobotanic indicators due to universal interrelations between natural components such as hard rocks, sediments, topography, waters, soils, and vegetation.

ACKNOWLEDGEMENTS

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STRUCTURAL CHARACTERISTICS OF PLEISTOCENE DEPOSITS IN NORTH LITHUANIA

by
Danguolė Karmazienė[1], Bronislavas Karmaza[2] and Valentinas Baltrūnas[2]

Karmazienė, D., Karmaza, B. & Baltrūnas, V. 2007. Structural characteristics of Pleistocene deposits in North Lithuania. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 31–38, 6 figures, 1 table.

Pleistocene till and intertill deposits were studied in the Upper Devonian dolomite of Petrašiūnai and Klovainiai (Pakruojis district) quarries in northern Lithuania. The sections were described in detail, the orientation and inclination of the long axes of gravel and pebbles were measured, and the samples for petrographic and grain size analysis were taken. The grain size of intertill silt as well as spore and pollen composition was determined. The dating was done using the OSL method. The thickness of the deposits studied in the Petrašiūnai area is up to 16 m and this sediment sequence was subdivided into three complexes. The Middle Pleistocene subglacial complex was formed by a glacier, which accumulated non-uniformly mixed local (Upper Devonian) clastic material and material originally from more remote northwestern areas. Basin sediments aged 86–81 thousand years (by OSL) are of special interest. They are underlying a glacial complex of the Baltija stage of the Nemunas (Weichselian) Ice Age. The complex is composed of tills accumulated in two phases by a glacier that was advancing from the north. The complex of tills differ from older tills because they have a lower content of imported clastic material and a higher content of local material.

Key words (GeoRef Thesaurus, AGI): glacial features, sediments, till, stratigraphy, grain size, Pleistocene, Lithuania.

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INTRODUCTION

The North Lithuanian region has been known as a zone of dominant glacial erosion where the Quaternary sediment cover is relatively thin (up to 40 m in palaeoincisions) and has a specific structure (Bucevičiūtė et al. 1992, and others). Previous research has shown that the character of glacial erosion and glacial material transportation are reflected in the structure of Pleistocene deposits, particularly in tills. The aim of the present paper is to report new information on structural peculiarities of North Lithuanian Pleistocene strata. The new data were obtained using different methods of studying Pleistocene deposits in the adjacent dolomite quarries near the Petrašiūnai and the Klovainiai boroughs in the Pakruojis District (Fig. 1). These quarries are 7–8 km away from each other. About 5–6 km north of the Petrašiūnai quarry, is the boundary of the North Lithuanian Phase of the Baltic Stage and is marked by a distinct morainic Linkuva ridge.

Fig. 1. Location of geological cross-section studied.

METHODS

Macroscopic description, graphic fixation and photography of the quarry walls were used to examine the Pleistocene succession the study site. Granulometric analysis of intertill silt served as a basis for stratigraphic classification of the Pleistocene strata. The following fractions were determined: 5–2; 2–1; 1–0.5; 0.5–0.25; 0.25–0.1; 0.1–0.05; 0.05–0.01; 0.01–0.005; 0.005–0.002; 0.002–0.001 and <0.001 mm. Spores and pollen analysis of samples taken from the quarry wall every 10 cm was carried out at the laboratory of the Lithuanian Geological Survey. The absolute age of these deposits was determined by the optically stimulated luminescence (OSL) method at the Laboratory of Quaternary Geochronology of the Institute of Geology, Tallinn Technical University. Recently published results on the granulometric, chemical and petrographic composition of tills were used for stratigraphic identification (Baltrūnas et al. 2005). A. Gaigalas' method was used for determining the petrographic composition, orientation and inclination of the long axes of gravel and pebbles (Ø 5-30 mm) in till units (Gaigalas 1971).

THE STRUCTURE OF THE PLEISTOCENE DEPOSITS OF THE REGION

The Pleistocene sediments in the central part of northern Lithuania are overlying the Upper Devonian dolomite, marl, limestone, clay, and gypsum layers. The varying resistance of these deposits to mechanical and chemical impacts as well as the different bedding and tectonic conditions were responsible for the different scale of erosion in various areas of the region. According to the available data, the palaeorelief is characterized by uplifts, solitary closed dips and troughs, and long and narrow palaeoincisions (Šliaupa 1997, 2004). A trough-shaped meridional declension whose axis extends across Vaškai towards Panevėžys stands out as a prominent feature. The minutely investigated sections of the Petrašiūnai and Klovainiai quarries are on the western slope of the declension. The sub-latitudinal Palaeomūša incision is situated north of the Petrašiūnai and Klovainiai sections.

According to the available borehole data, the Pleistocene deposits are composed of three complexes of glaciogenic sediments (Fig. 2). In the stratigraphic scheme used by the Lithuanian Geological Survey, these glacial deposits are attributed to the Middle Pleistocene Medininkai and the Upper Pleistocene Upper Nemunas ice ages (sub-formations) (Lietuvos … 1994). The Pleistocene strata are rich in intertill deposits. The glacial deposits of the Medininkai sub-formation are spread locally in the region, overlie Upper Devonian rocks and are represented mainly by brown sandy loam. The till under consideration is very compact. Its greatest thickness (38 m) was measured in a palaeoincision 24 km NW of Pakruojis, with the prevailing thickness being 10–15 m. In some places it is only 3–5 m. The till left by the Medininkai glacier includes some blocks of Devonian rocks. The thickest block (21 m) was found in a borehole 28 km north of Pakruojis.

Glacial deposits of the Upper Nemunas Formation cover the whole territory. These deposits overlie the Medininkai Formation or, in rarer cases, Upper Devonian rocks. At some places, the glacial depos-

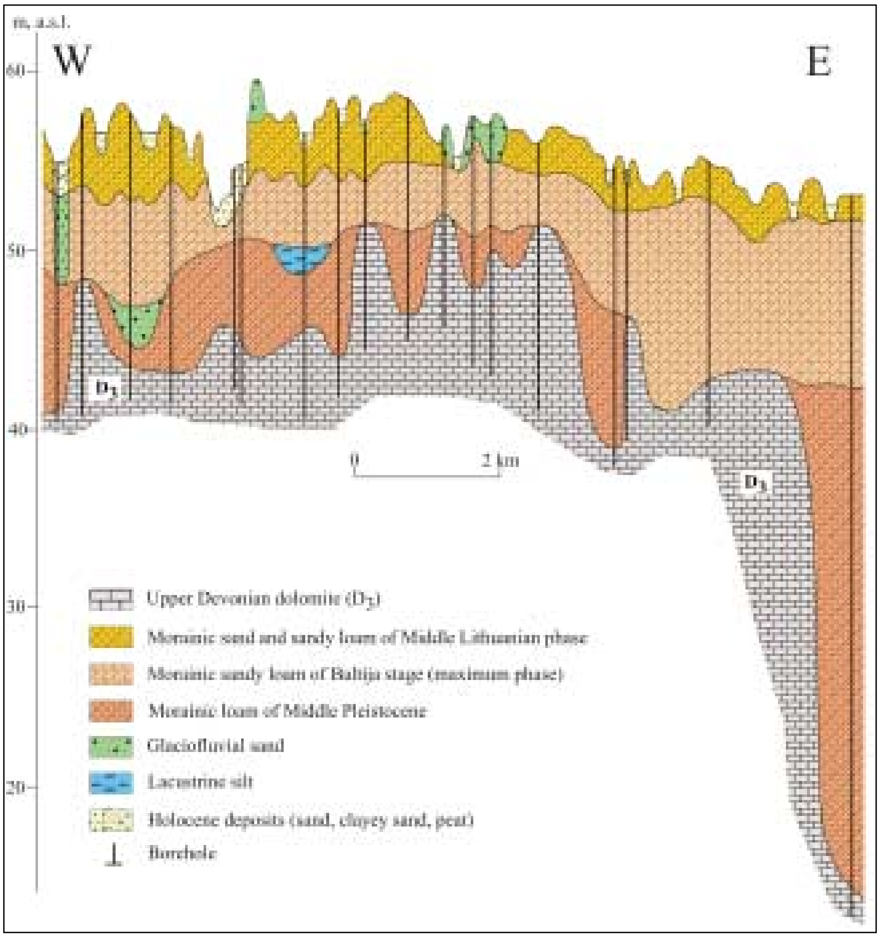


Fig. 2. Cross-section of the Petrašiūnai area Pleistocene deposits.

its of the Upper Nemunas underlie glaciofluvial, lacustrine or other types of sediments. The greatest thickness of this formation (34 m) was measured in an incision near the Žeimelis borough. The deposits are mostly represented by greyish brown or brown sandy loam.

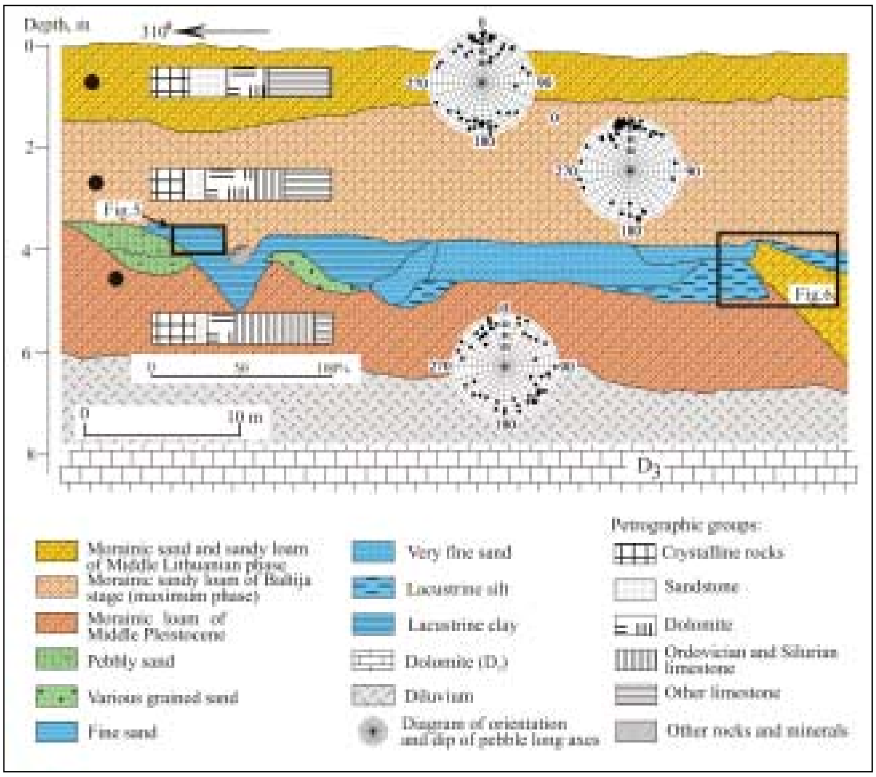
THE STRUCTURE OF PLEISTOCENE SEDIMENTS IN THE PETRAŠIŪNAI AREA

The pre-Quaternary surface in this area is composed of the dolomite of the Upper Devonian Stipinai Formation with interlayers of marl and clay. Its absolute altitude ranges between 39 and 58 m a.s.l. The maximum thickness of the Pleistocene deposits (14 m) was measured in the western part of the area, and the minimum thickness (1-2 m) was found in the eastern part. The thickness of Pleistocene deposits is inversely proportional to the altitude of pre-Quaternary rocks (the correlation coefficient -0.81). The greatest thickness of these deposits is usually measured in the surface depressions of Devonian rocks.

The lower (older) till complex has a triple structure. In the lower part, it is composed of brown clayey solid till with a poorly expressed orientation of the long axes of gravel and pebbles. In the middle part of the complex (up to 1.5 m thick), brown clayey till passes into grey less clayey till with a greater mixture of gravel (Fig. 3). In the upper part of the complex, grey till (characterized by its mostly NW–SE orientation of the long axes of gravel and pebbles) merges into similar brown till. According to research data on intertill deposits, this till complex is preliminarily attributed to the Middle Pleistocene Medininkai sub-formation.

The intertill complex in the surroundings of the Petrašiūnai quarry is widespread, lithologically diverse but thin (up to 2 m) (Fig. 4). Yellowish grey and greyish yellow silt, sandy silt and fine-grained silt sand in places underlie a layer of varves up to 1 m in thickness (Fig. 5). Occasionally, the varves are disordered and have patches of gravel and pebbles. The brown and dark brown solid till in the Petrašiūnai-1 section (north western wall of the Petrašiūnai quarry) underlies compact, subhorizontally stratified grey silt 0.45 m thick, which is greyish brown in the lower part. The grey silt underlies a 0.8 m thick layer of brownish grey slanted, finely stratified silt with a greater mixture of sand (especially in the middle part). The upper part of

Fig. 3. Cross-section of Petrašiūnai-2 Pleistocene deposits in the southwestern part of the dolomite quarry (24-08-2002).



the complex is composed of 0.50 m thick greyish yellow indistinctly horizontal stratified silt with an even greater content of sand.

Pollen analysis of the 1.75 m thick silt layer has shown that pollen is well preserved. Its concentration is small. At the time of silt and sandy silt deposition, the climate was cold and the plant cover poor. The spores and pollen spectra provided no basis for a more precise biostratigraphcal correlation. According to the OSL dating results, the deposition took place 86.1–81.6 thousand years ago (Table 1).

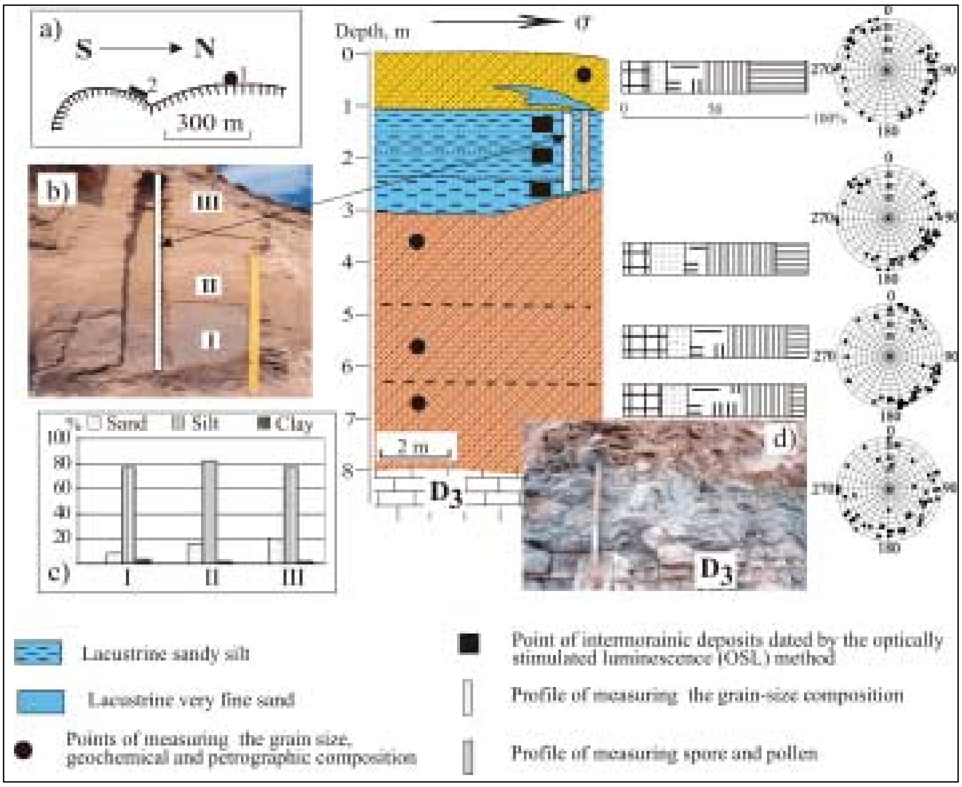


Fig. 4. Cross-section of Petrašiūnai-1 Pleistocene deposits in the southwestern part of the dolomite quarry (24-08-2002): a – location of Petrašiūnai-1 and Petrašiūnai-2 cross-sections; b – photograph of intermorainic sediments; c – grain-size composition in separate parts (I, II, III) of intermorainic sediments; d – photograph of Devonian dolomite contact with till. Other explanations as in Fig.3.

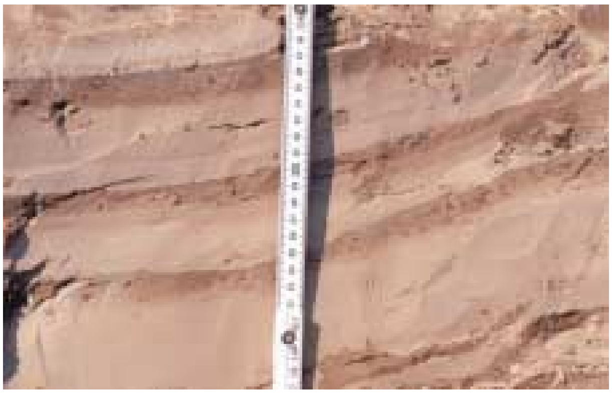


Fig. 5. Photograph of the intermorainic varved clay.

Table 1. Results of OSL dating of intertill sediments in Petrašiūnai-1 section.

Sample No	Depth, m	Laboratory code	U (ppm)	Th (ppm)	K (%)	$D_{ext\ sed}$ (μGy/a)	P_s (Gy)	OSL-age, ka
1	1.3–1.4	TLN 1507-093	1.00	3.70	1.76	2150	232.50	81.6±6.5
2	1.7–1.8	TLN 1508-093	1.20	3.36	1.77	2185	242.00	84.4±13.7
3	2.3–2.4	TLN 1509-093	0.99	3.92	1.93	2212	248.00	86.1±9.3

The upper (Nemunas Glacial of the Baltic Stage) till complex can be divided into two characteristic units. The thickness of the brown clayey till unit is up to 3 m (Fig. 3). Somewhere, together with intertill varves, it occurs at a lower hypsometric level. The petrographic composition of gravel and pebbles in till is characterized by the prevalence of local clastic material (Devonian dolomites and limestones) and a smaller portion of imported clastic material (Ordovician and Silurian limestones). The long axes of gravel and pebbles are oriented N–S.

The second surface layer is composed of light brown, pinkish brown and brownish yellow sandy loam and till of the Middle Lithuanian Phase of the Baltic Stage. Its thickness reaches 1.5 m. The clayey portion (<0.01 mm) accounts for only 6.7-8.3 %. The till is massive and friable, with prevailing pebbles and typical weathered gneisses. At the base, sand prevails. The lower margin is distinct, with oblong patches of very fine sand and silt. In comparison to the older tills, the surface till contains less imported Ordovician and Silurian limestones and crystalline rocks.

FORMATION FEATURES OF PLEISTOCENE SEDIMENTS: DISCUSSION

According to the presented material, the greatest part of North Lithuanian Pleistocene sediments is composed of glacial deposits. The structure of these sediments, the orientation and inclination of the long axes of gravel and pebbles in till and sometimes the variation of the index of relative entropy of granulometric composition best reflects the dynamic conditions of sedimentation. The lower (older) till of the Petrašiūnai quarry is heterogeneous. The dominant orientation of the long axes of clastic material is NW–SE, with the typical inclination being SE. The typical orientation and inclination are absent only in the lower part of the till. The relative entropy is low, showing a poor mixing of till material (fraction <0.01 mm accounts for 17.3-18.3%) (Baltrūnas et al. 2005). The orientation of the long axes of gravel and pebbles and the petrographic composition of tills are not typical of the Medininkai sub-formation tills. They are comparable with the Nemunas upper sub-formation Grūda and the Nemunas lower sub-formation Varduva tills. The age of the latter till corresponds with the age of the overlying intertill sediments determined by the OSL method. The determined age and cool palaeoclimatic conditions correspond to the marine oxygen isotope stage 5a of the Upper Pleistocene, which can be correlated in Lithuania with the end of the Merkinė (Eemian) Interglacial (Gaigalas 2001) or with the Early Nemunas warming (Jonionys 2) and in the Central Europe, with the Odderade Interstadial (Satkūnas 1999).

The till of the maximal phase of the Baltic Stage is distinguished for a high content of clay (fraction <0.01 mm accounts for 25%), high relative entropy of granulometric composition and the dominant N–S orientation and N inclination of the long axes of gravel and pebbles in till. The till and sandy loam of the Middle Lithuanian phase in the Petrašiūnai area is characterized by a small relative entropy (poor sorting) of its granulometric composition and a well-expressed orientation and inclination of the long axes of gravel and pebbles in tills. Till in the lower part of the Petrašiūnai quarry is rich in the underlying local Devonian rock material (especially in yellowish and greenish grey dolomite with marl and clay interlayers). It also contains a large block of older till (Fig. 6). The implication is that these tills in the Petrašiūnai area were formed not only from imported clastic material (crystalline rocks, Ordovician and Silurian limestones), but also from excavated local older deposits.


Fig. 6. Photograph of dislocated lower till.

CONCLUSIONS

The thin (up to 30-40 m) Pleistocene bed in the North Lithuanian region is subdivided into three lithostratigraphic complexes. These are: the lower glacial complex of the Middle Pleistocene (Medininkai sub-formation) or Upper Pleistocene (Lower Nemunas sub-formation), the overlying intertill sediments aged 86-81 thousand years and the upper till complex of the Baltic Stage of the last Ice Age.

The lower (older) glacial complex in the pre-Quaternary declensions was formed in the interface between a glacier advancing from the north and a pre-Quaternary surface inclined to the north and the east.

The Upper Pleistocene glacial complex of the Baltic Stage of the last Ice Age is composed of strata accumulated during two recessions of the glacier that advanced from the north. The strata differ from the older tills in a lower content of imported clastic material and a higher content of local excavated material. The strata differ among themselves in granulometric composition and sorting.

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PERIGLACIAL CONDITIONS AND DEGLACIATION IN SOUTHERN LITHUANIA DURING THE LAST ICE AGE

by
Valentinas Baltrūnas[1,2], Kęstutis Švedas[2] and Violeta Pukelytė[1]

Baltrūnas, V., Švedas, K. & Pukelytė, V. 2007. Periglacial conditions and deglaciation in Southern Lithuania during the Last Ice Age. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 39–46, 4 figures, 2 tables.

The southern Lithuanian territory is composed of glaciogenic deposits of different age and polygenetic origin. The recurring permafrost and cryogenic structures in the ground were an important phenomenon taking place in the southeastern periglacial zone. Intensive slope transformation took place together with permafrost degradation and deposition of the loess-like sediment cover. These events were followed by soil formation processes. Sediment cover is composed of three to four lithocomplexes that correlate with palaeogeographic stages of the Last Ice Age. The boundary of the Grūda (Brandenburg) stage glaciation correlates with the maximum of the Late Nemunas (Late Weichselian) and is conditionally drawn along the poorly preserved and sporadically distributed abraded hills. The marginal hilly formations, left by the glacier of the Žiogeliai (Frankfurt) phase, are clearly marked by the high moraine ridge on the right bank of Merkys. The specific character of Baltija (Pomeranian) deglaciation was predetermined by the wide (30-40 m), dead ice elevation and the Simnas-Balbieriškis-Stakliškės glaciolacustrine basin, dammed between the South Lithuania Upland and the younger South Lithuanian glacier. Its excess water eroded a great part of the Middle Nemunas valley between Punia and Merkinė in a southerly direction. The middle part of the territory, extending from the southwest to the northeast is a part of Vilnius- Warsaw-Berlin Urstromtal.

Key words (GeoRef Thesaurus, AGI): glacial geology, paleogeography, periglacial environment, deglaciation, glacial features, sediments, Pleistocene, Lithuania.

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INTRODUCTION

Many publications have been dedicated to issues of the Late Pleistocene palaeogeography in Lithuania. Some of them are theoretical in nature and are confined to general and schematic palaeogeographical characteristics of the region. The publications served both as a theoretical and methodological basis for more recent investigations (Basalykas 1965, Dvareckas 1993). Some other studies were more specific with a special method used for analyzing diverse geologic and geomorphologic objects and with fundamental conclusions substantially contributing to the knowledge of the history of palaeogeographic conditions of the time (Basalykas et al. 1984, Dvareckas & Dicevičienė 1987, Kabailienė 1990, Kondratienė 1996, Komarovsky 1996, Satkūnas 1999, Gaigalas 2001, Švedas 2001, Švedas et al. 2004). Furthermore, it should be recognized that a composite evaluation and more accurate cartographic depiction of the palaeogeographical conditions were not fully realized, and the correlation of Pleistocene stratigraphical events in Lithuania was not associated with changes in the periglacial conditions.

The purpose of this article is to summarise previous investigations on the geology and palaeogeography of the last glaciation and soil permafrost structures in periglacial areas in southern Lithuania to obtain a comprehensive model of the evolution of the area (Fig. 1).

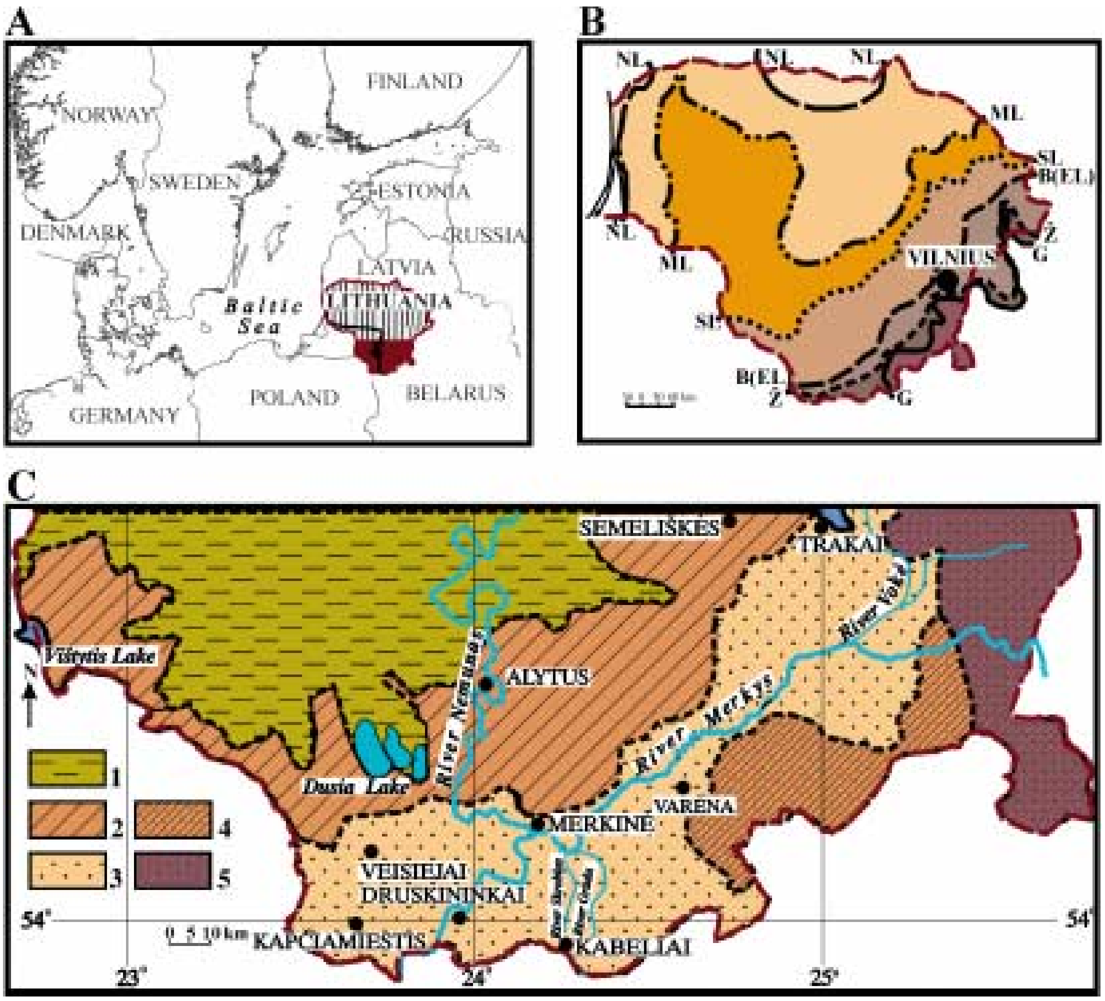

Fig. 1a) Location map of the studied area, b) Deglaciation of the Last (Weichselian, Nemunas) Glaciation in the South Lithuania (after Gaigalas (2001)): G – Grūda (Brandenburg) stage, Ž – Žiogeliai (Frankfurt) phase, B (EL) – Baltija (Pomeranian) stage (East-Lithuanian phase), SL – South-Lithuanian phase, ML – Middle-Lithuanian phase, NL – North-Lithuanian phase; c) Geomorphological regions (compiled according to Atlas of Lithuania (Drobnys 1981): 1 – Middle Nemunas Plateau, 2 – South Lithuania Upland, 3 – Dainava Plain, 4 – Eišiškės Plateau, 5 – Medininkai Upland.

METHODS

Pleistocene deposits in southern Lithuania have been identified and described using granulometric, mineralogical (in fractions 0.2–0.1 and 0.1–0.05 mm), petrographic (fractions 30–10, 10–5, and 5–2 mm), and geochemical (fraction <1 mm) analyses. Interglacial and post-glacial sediments were identified and spore, pollen, and diatomic analyses included carbon dating (^{14}C). Analysis results were presented by Baltrūnas (2001, 2002), Gaigalas (2001), Kabailienė (2001) and others.

Examination of buried soils helped to determine the stratification of sediments and reconstruct the palaeogeographic history. The content of organic matter (Corg), acidity of the environment (pH), concentrations of iron (Fe, Fe^{+2}, Fe^{+3}) and carbonates ($CaCO_3$ and $CaMg(CO_3)_2$) as well as the composition of clay minerals provided information about the geochemical environment. The layers of buried soils and cryogenic structures (pseudomorphoses and involutions) disturbing the sediments were helpful in distinguishing the lithocomplexes of sediments of different ages and estimating the time of their formation (Švedas 2001). Permafrost-induced structures of the soil, and a systematic classification of glacial and periglacial environments and deposits have been dealt with in a special work (Brodzikowski & Van Loon 1991). The palaeogeographic schemes were compiled using the original method and legend (Baltrūnas 2002, Švedas et al. 2004).

STRUCTURE OF SEDIMENT COVER IN THE ZONE OF PENULTIMATE GLACIATION

The petrographic analysis of gravel and pebbles in till shows that the Medininkai Upland and Eišiškės Plateau have similar composition, but considerably differ from the South Lithuania Upland, particularly from the part formed during the Baltija (Pomeranian) stage (Table 1). This is confirmed by the ratio of petrographic coefficients: D/L (dolomite/Ordovician and Silurian limestone) and C/S (total of crystalline rocks/total of sedimentary rocks). Comparison of the studied areas according to the composition of heavy minerals (fraction 0.2-0.1 mm) revealed how similar the Medininkai Upland and Eišiškės Plateau are and how they differ from the South Lithuanian Upland (Table 2). The values of the coefficient K_m, obtained from the ratio of the total sum of "Fennoscandian" (hornblende and pyroxenes) and "local" (iron oxides and hydroxides, zircon, epidote and apatite) minerals, also indicate the similarity between the two areas. The sedimentation environments and post-sedimentation processes, particularly the periglacial conditions, determine the composition of glacigenic deposits.

A thick (5–7 m) layer of sandy loam and clayey material covers the glaciogenic surface of the Medininkai Upland. During detailed field and laboratory investigations, three (in some places even four) lithocomplexes were distinguished (Fig. 2). The distinguished lithocomplexes differ in composition, colour, accumulations of iron and carbonates, and the cryogenic structures in them. Also, these lithocomplexes are separated by thin layers of organic matter. However, the content of C_{org} in them is not high – 0.28–0.70%. These layers correspond with hydromorphic, poorly developed soils. Similar buried soils in the extraglacial zones of Poland have been described by Polish and Belarus researchers (Jary et al. 2002, Komarovsky 1996).

The lower (1-3 m) lithocomplex is composed of unstratified fine-grained sand and sandy loam with clay and gravel beds as a result of the process of

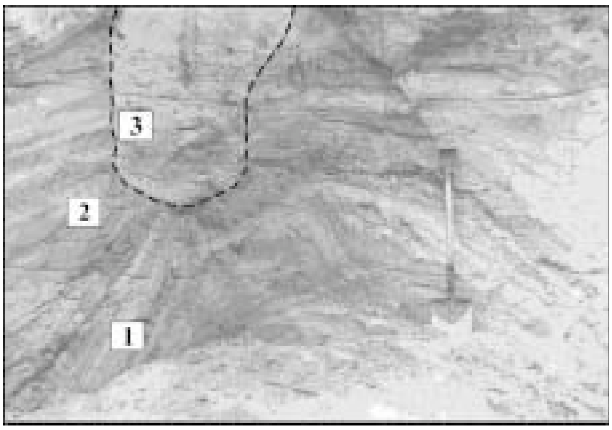

Fig. 2. Structure of sediment cover in Eišiškės Plateau (Kalesninkai environs): 1 – lower lithocomplex, 2 – intermediate lithocomplex, 3 – upper lithocomplex.

segregation. A great variability of lithofacies and involutions of cryogenic origin are characteristic of this lithocomplex.

The intermediate lithocomplex (about 2 m in thickness) is also composed of fine-grained sandy loam-like material. It abounds in stripes of iron oxide accumulations, involution structures, and different size of pseudomorphoses, formed in places of former ice wedges.

The upper lithocomplex is composed of unstructured fine-grained sands, sandy loam and silt with sparse gravel beds with a thickness of 1–2 m. The involutions are not very distinct and the pseudomorphoses are sparse. The upper lithocomplex is covered by the modern soil.

The high amount of fine-grained material in the lithocomplexes (fraction 0.1–0.05 mm – 48.4%, 0.05–0.01 mm – 23.5%, clay particles (< 0.01 mm) – from 9.8% to 18.4%) implies a strong influence of cryogenic processes on the soils. A recurring cycle of freezing and thawing accelerates the rock weathering rates. As a result, the silt and clay particles accumulate at high rates, and the soils become similar to loess.

Table 1. Arithmetical means (X, %), variation coefficients (v, %) and petrographic coefficients (D/L, S/C) of gravel and pebble petrographic groups of surface till in southern Lithuania. Petrographic coefficients: D/L – dolomite/Ordovician and Silurian limestone; C/S – total of crystalline rocks/total of sedimentary rocks.

Petrographic groups		**Geomorphological regions**			
		Medininkai Upland	**Eišiškės Plateau**	**Žiogeliai zone of South Lithuania Upland**	**Baltija zone of South Lithuania Upland**
Number of samples		15	19	15	15
Quartz	X v	4.4 47	6.4 50	2.2 54	3.6 122
Potassium feldspars	X v	3.7 67	4.3 79	2.2 76	2.2 99
Crystalline rocks	X v	54.6 43	51.4 38	39.7 36	37.2 39
Sandstone	X v	8.2 49	11.2 54	11.7 36	9.6 46
Dolomite	X v	5.6 112	7.8 112	5.1 62	9.4 70
Ordovician and Silurian limestone	X v	1.9 203	2.7 213	8.4 72	11.5 87
Other limestone	X v	20.0 94	13.2 113	28.1 53	23.9 61
Petrographic coefficients	D/L C/S	2.95 1.76	2.89 1.78	0.61 0.79	0.83 0.82

Table 2. Arithmetical means (X, %), variation coefficients (v, %) and mineralogical coefficient (Km) of some heavy minerals (fraction 0.2-0.1 mm) of surface till in South Lithuania. Km - hornblende and pyroxenes / iron oxides and hydroxides, zircon, epidote, apatite.

Heavy minerals		Geomorphological regions			
		Medininkai Upland	**Eišiškės Plateau**	**Žiogeliai zone of South Lithuania Upland**	**Baltija zone of South Lithuania Upland**
Number of samples		15	19	15	15
Iron oxides and hydroxides	X v	5.6 46	6.9 21	5.3 17	5.6 20
Magnetite and Ilmenite	X v	14.8 41	16.6 27	17.8 9	18.8 9
Zircon	X v	2.2 52	2.5 48	2.2 30	1.9 40
Garnet	X v	22.8 41	23.9 25	27.1 7	26.8 5
Apatite	X v	2.3 43	2.2 44	1.8 33	1.6 35
Epidote	X v	5.1 45	5.6 33	6.6 14	6.1 13
Hornblende	X v	25.6 41	28.0 26	28.3 5	28.2 5
Pyroxene	X v	2.7 51	2.9 32	5.0 21	4.3 2.14
K_m		1.86	1.80	2.09	2.14

CHANGE OF PALAEOGEOGRAPHICAL CONDITIONS IN THE TERRITORY OF PENULTIMATE GLACIATION

The damp and warm interglacial climate was followed by a period of soil degradation, associated with the cold climate at the beginning of the Weichselian Glaciation. The ground became deeply frozen and permafrost fissures appeared. The soils of that time contained pseudomorphoses, involutions and other kinds of cryoturbations as well as coarse-grained sand and gravel, accumulated during segregation in permafrost fissures. The eolian processes then reworked the fine-grained silt in the lower lithocomplex. These events took place in the Early Weichselian.

After the long-lasting cold climate, a moderately warm climate set in. There appeared rudiments of soil forming processes in coniferous forests. Traces of these processes were identified under the cover of fine-grained sand and silt. The following cooling of the climate disturbed the deposited sediments by new cryoturbations. The material from that time is saturated with iron oxides. It also contains carbonate accumulations and poorly expressed pseudomorphoses. The formation of this layer is associated with the Middle Weichselian. Fine-grained loess-like material was deposited at the end of this period. The dry and cool climate followed at the end of Middle Weichselian causing a long-lasting permafrost and degradation of soil horizon. The sediments included stratified solifluction matter.

The last phase of soil formation is related to the moderately cold climate of the Upper Pleniglacial. The formation of pseudogleyic soils resumed. When the climate cooled in its final stage, the layer of frozen ground thickened considerably, the weathering processes intensified, and the layer of the upper lithocomplex was formed. It also contains well expressed marks of cryoturbation. With the climate warming and permafrost degradation at the end of this period, many disturbed soils appeared, indicating the solifluction intensity.

PALAEOGEOGRAPHY OF SOUTH LITHUANIA IN THE ZONE OF LAST GLACIATION

During the Late Weichselian, a large part of southern Lithuania was covered by the glacier advancing from northwest. During the Grūda stage (Brandenburg, Late Weichselian) its edge extended along the western slope bottom of the Medininkai Upland and only partly covered the western part of the Eišiškės Plateau. Poorly expressed end moraines have been detected. The edge of the glacier (presumably not very thick) is marked by low washed out sand and sandy loam hills.

The degradation of the last glacier of the maximal Grūda (Branderburg, Late Weichselian) stage produced a huge amount of melt water dammed in between the decaying glacier and the southernmost moraine hills. Thus, the periglacial lakes were formed with water levels at first between 180–160 m a.s.l. Later, the Verseka, Ditva, and Juodupė valleys between the ridges became the outlets for southern water flows.

The new distinct Žiogeliai (Frankfurt) phase of glaciation left well developed marginal moraine formations on the left bank of the Merkys (Fig. 3) that are overlying the glaciolacustrine fine- and very fine-grained sand of the Grūda (Brandenburg) stage. The formation of glaciofluvial (sandur and sandur deltas) sediments was closely linked with the history of glaciolacustrine basins. This assumption is supported by the absolute surface altitude of the mentioned sediments, which gradually decreases from 160 m a.s.l. at the northeastern edge of the glacier to 130–125 m a.s.l. at the southwestern edge. This also implies an asynchronous character of their formation. It is possible to assert that the glacier edge stabilization at Kapčamiestis, Liškiava, Onuškis, and Trakai in the Žiogeliai (Frankfurt) phase was related to a short climate warming that slowed down solifluction and renewed cryogenic processes in the Medininkai Upland and Eišiškės Plateau.

The Baltija (Pomeranian) palaeogeography of the studied region was associated with the Middle Nemunas ice lobe that advanced from the north, which was unable to overcome the characteristic high moraine ridge of Žiogeliai phase. During the glacier advance and stabilization at Veisiejai, Merkinė, and Semeliškės, the previously formed relief endured cryogenic processes.

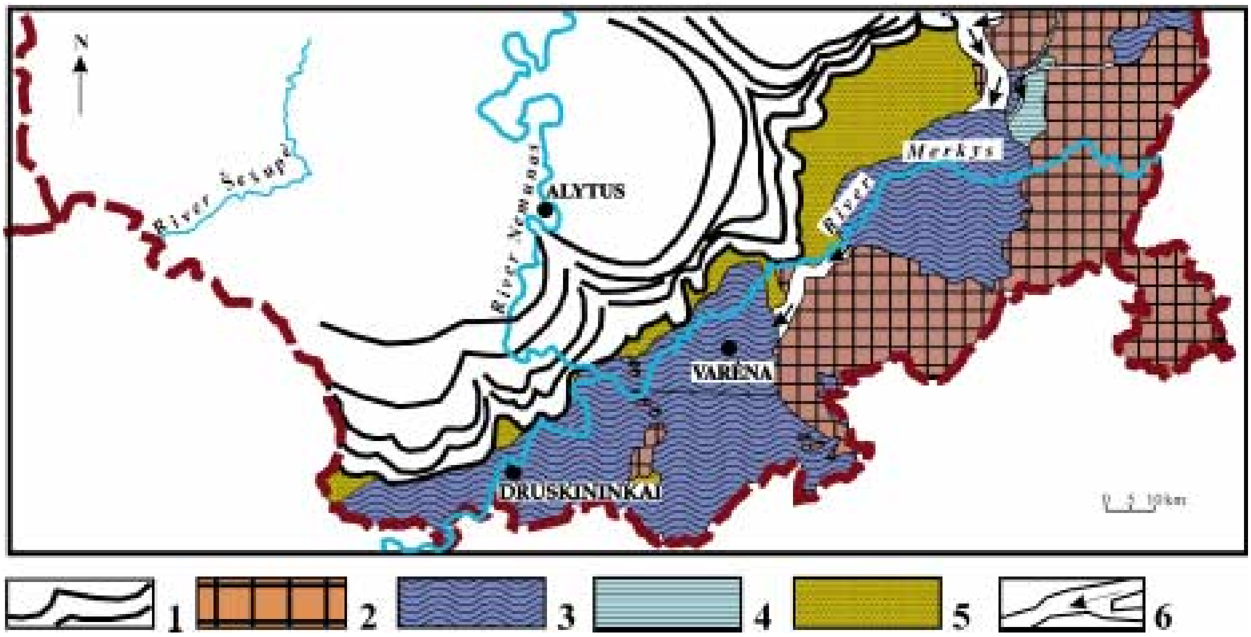

Fig. 3. Palaeogeographical situation of the Žiogeliai (Frankfurt) phase of Nemunas (Weichselian) Glaciation in southern Lithuania: 1 – active glacier; 2 – hilly relief marginal deposits of next-to-last (Medininkai, Saalian) glaciation formed cryogenic and solifluction processes; 3 – glaciolacustrine basin; 4 – glaciolacustrine plain; 5 – the formed sandur (glaciofluvial) plain; 6 – valley and direction of water flush.

The South Lithuanian phase of glacier degradation was marked by a recessive pause north of Virbalis and Marijampolė at Igliauka and Prienai (Fig. 4). A lake valley, drained by melt water and opening into the large Simnas–Stakliškės glaciolacustrine basin existed between the glacier and the upland, which in many places was still covered by thawing dead ice with moraine sediments. The South Lithuanian Upland in the southern and eastern coasts of this basin was also covered by moraine-bearing dead ice. A wide southward valley between high hills served as a passage for the excess water of the basin. The Merkinė–Punia sector of the Middle Nemunas River valley, formed by shallow water streams flowing from the northern to the southern glaciolacustrine basins, began to form at exactly that time. In the initial stage, the water level in the"overflowing" basins was 125 and 120 m a.s.l. respectively. Later, the level of the northern basin fell to 115–110 m a.s.l. and the Nemunas River reversed the current direction. The level of the southern basin was more stable and related to the stable and intensive melt water flow through the lateral old valley from eastern Lithuania (Vilnius–Warsaw–Berlin Urstromtal) (Basalykas et al. 1984). The retreat of glaciers from southern Lithuania and the outflow of periglacial basins formed the geological and geomorphological basis of the region. Later, this basis was transformed by various other geological processes (solifluction, glaciokarst, aeolian, and erosion).

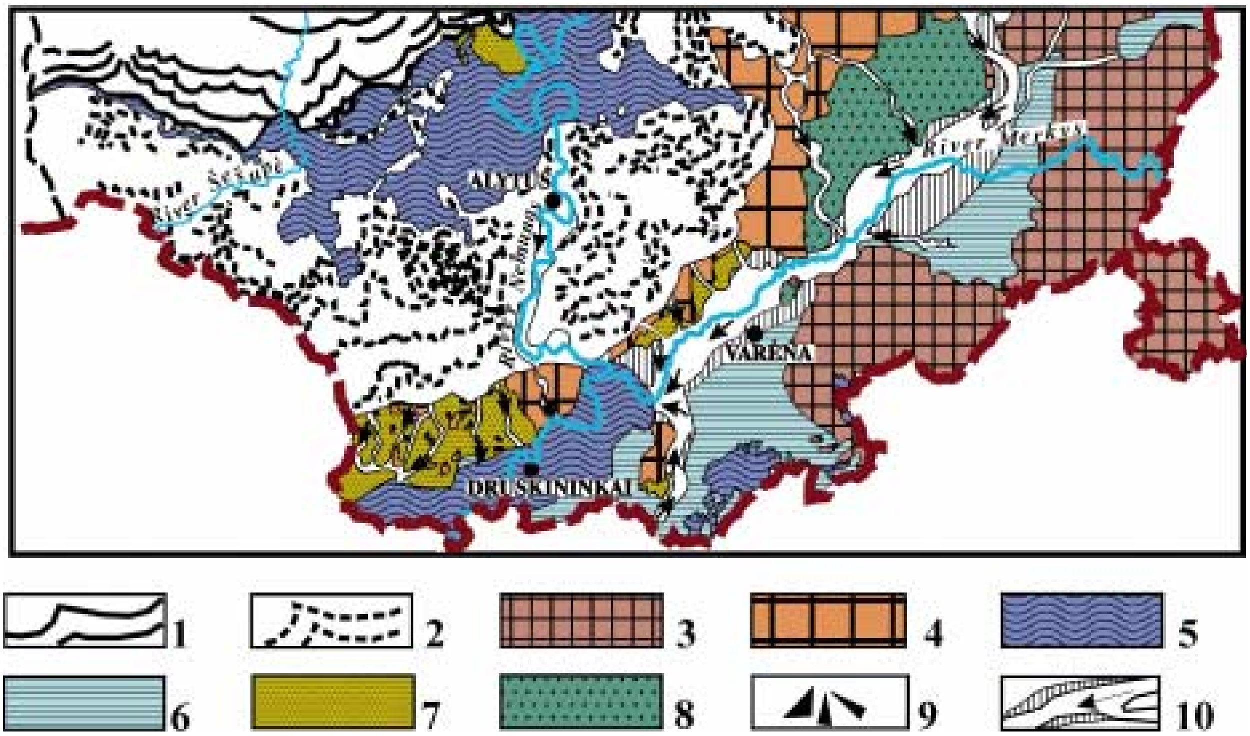

Fig. 4. Palaeogeographical situation of South–Lithuanian phase of the Nemunas (Weichselian) Glaciation in southern Lithuania: 1 – active glacier; 2 – dead and thawing glacier; 3 – marginal deposits in hilly relief of next-to-last (Medininkai, Saalian) glaciation formed by cryogenic and solifluction processes; 4 – marginal deposits of hilly relief of Last (Nemunas, Weichselian) Glaciation; 5 – glaciolacustrine basin; 6 – glaciolacustrine plain; 7 – the formed sandur (glaciofluvial) plain; 8 – sandur (glaciofluvial) plain; 9 – glaciofluvial delta; 10 – valley, terrace and direction of water flush.

CONCLUSIONS

1. The southeastern periphery of southern Lithuania endured long-lasting periglacial conditions. The northwestern part was impacted by the glaciogenic sedimentation of the Žiogeliai (Frankfurt) and Baltija (Pomeranian) stages of the Last (Weichselian) Glaciation. The middle part, extending from south-west to north-east was influenced by very intensive melt water erosion and accumulation (Vilnius–Warsaw–Berlin Urstromtal).
2. The permafrost and cryogenic structures represented an important phenomenon of extraglacial zones. Permafrost degradation consisted of intensive slope transformation processes, which influenced the formation of loess-like structures. The permafrost degradation began in southern Lithuania in conjunction with the melting of the Baltija (Pomeranian) glaciers about 16,500–13,500 years ago.
3. The specific features of the Baltija stage deglaciation were predetermined by a wide (30–40 m) elevation of dead ice and a glaciolacustrine basin (Simnas–Stakliškės) dammed between the the South Lithuania Upland and the younger South Lithuania glacier. The excess water of this basin washed out a large part of the Middle Nemunas River.

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Geological Survey of Finland, Special Paper 46, 47–54, 2007.

LATE WEICHSELIAN DEGLACIATION IN FINNISH LAPLAND

by
Peter Johansson[1]

Johansson, P. 2007. Late Weichselian deglaciation in Finnish Lapland. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 47–54, 4 figures.

The picture of the Late Weichselian deglaciation in northern Finland is based on studies of glacial processes and morphology. These include subglacial meltwater action and glaciofluvial hydrography, till stratigraphy and ice flow directions, proglacial, marginal and lateral meltwater action and development of ice lakes. The deglaciation started from the northernmost part of the area as the ice margin retreated from the Younger Dryas End Moraines about 11,600 years ago. The last remains of the glacier disappeared from western Finnish Lapland about 10,000 years ago. Using all the palaeohydrographic information, for example by mapping the ice lake phases and the routes of the extramarginal channels between them, it has been possible to form a reliable reconstruction of successive stages of the ice retreat during these 1,600 years.

Key words (GeoRef Thesaurus, AGI): glacial geology, deglaciation, glaciofluvial features, meltwater channels, glacial lakes, Pleistocene, Weichselian, Lapland Province, Finland.

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INTRODUCTION

The deglaciation phase and the final disappearance of the Late Weichselian ice sheet were examined in Finnish Lapland by studying different glacial processes and their resulting landforms. These are subglacial meltwater action and glaciofluvial hydrography, till stratigraphy and ice flow directions, proglacial, marginal and lateral meltwater action and the development and types of the ice lakes.

Tanner (1915) was the first to study deglaciation in northern Finland. The first model of the retreating glacial margin was created by him and is still valid in a broad scale. Mikkola (1932), Penttilä (1963), Kujansuu (1967), Johansson (1995) and many others have by their research added more details to the picture of the deglaciation. Compared with southern and central Finland, the advantage of studying deglaciation in northern Finland lies in the fact that most of the retreating ice sheet melted in a supra-aquatic environment, resulting in various kinds of erosional and depositional landforms. Subaquatic conditions existed only in the southwestern part of Lapland because that area was covered by the waters of the Ancylus Lake phase of the Baltic Basin. Some small subaquatic areas exist in the northern part, where the Arctic Ocean for a short time penetrated into the river valleys of Teno and Lutto (Nikonov 1964, Saarnisto 1973) and into the Inari Lake basin (Tanner 1915, 1936, Kujansuu et al. 1998).

YOUNGER DRYAS END MORAINE

The Younger Dryas ice-marginal landforms in northern Norway are situated only 20 kilometres from northern Finnish Lapland (Sollid et al. 1973, Andersen et al. 1995). The active ice flow stage of Younger Dryas can be detected in Inari, northern Lapland, which was considerably influenced by the distinct ice lobe during the Younger Dryas (Punkari 1980). The extensive drumlin field is related to this flow stage and it extends as far as the Younger Dryas ice-marginal landforms in Norway and Pechenga area, NW Russia (Marthinussen 1962, Yevzerov & Koshechkin 1980, Yevzerov & Kolka 1993, Yevzerov et al. 2005, Semenova 2005). During this time the so called younger till bed or Till Bed II was deposited and occurs widely through the area. On top of the Till Bed II there is a quite common even younger till bed, known as Till Bed I or surficial till (Hirvas 1991, Johansson 1995). Its deposition is related to the ice flow during the deglaciation following the Younger Dryas. In eastern Lapland the youngest ice flow direction can be used to picture the retreat of the ice sheet, because the retreat was usually in the opposite direction to that of the last ice flow. The network of subglacial glaciofluvial systems indicates the direction of the retreating ice sheet even more accurately than the till fabric of the surficial till. This is due to the fact that in subglacial tunnels the meltwater tends to flow disregarding any terrain obstacles toward the lowest pressure i.e. the margin of the glacier. The subglacial meltwater action was mainly erosional at first, forming channels and gorges along the path of the conduit. Later, deposition prevailed closer to the ice margin and it was then that steep-sided and sharp-crested esker ridges were formed (Fig. 1). The distribution pattern of the Late Weichselian subglacial meltwater system is radial (Fig. 2) and suggests that the direction of the ice retreat was from the Younger Dryas End Moraines in Norway and Russia towards the ice divide area in northern central Lapland.

Fig. 1. The Pakajärvi esker is a typical subglacial esker ridge with a sharp crest and steep sides.

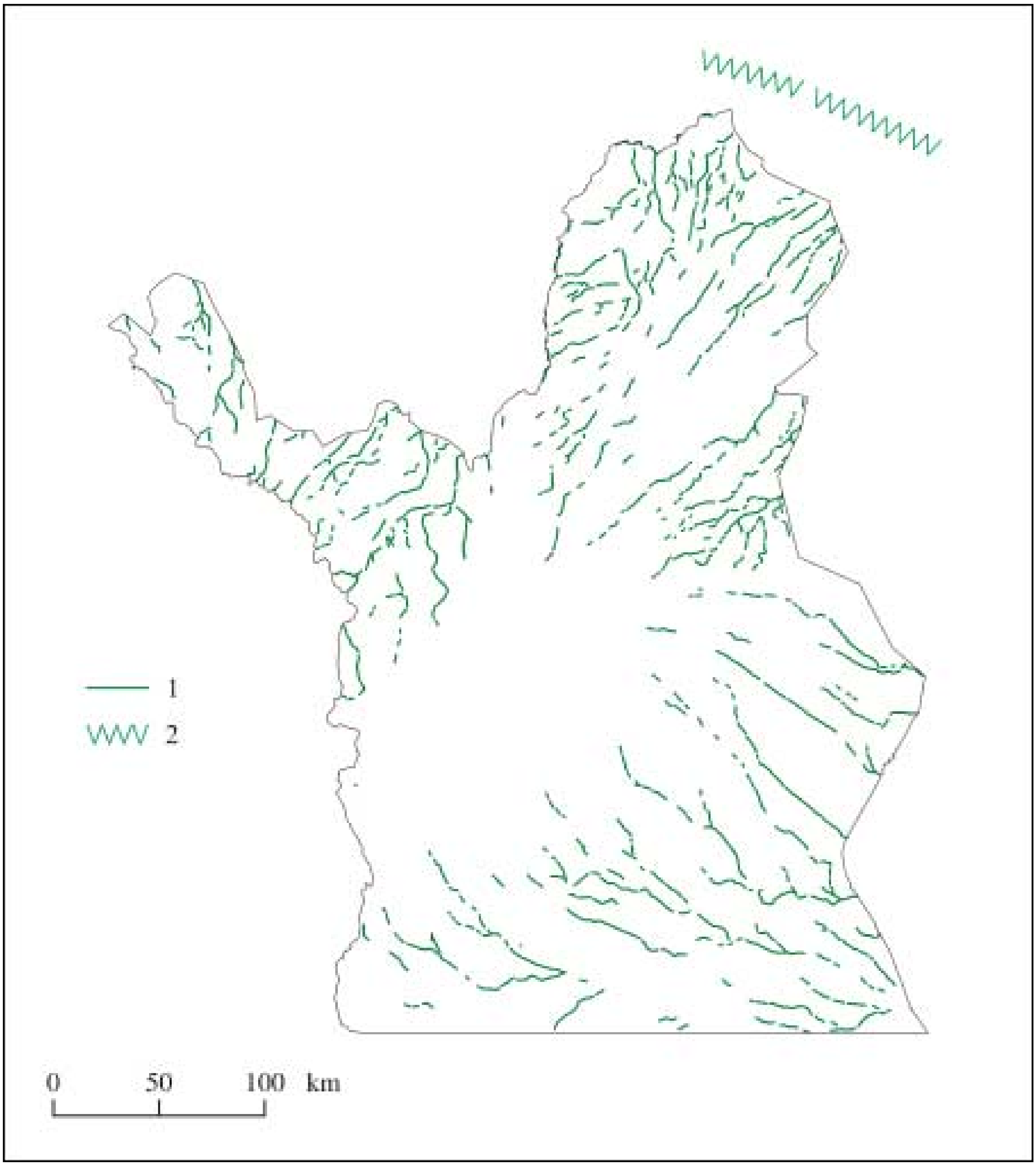

Fig. 2. Late Weichselian esker chains form a radial system which represent the direction of ice retreat from the Younger Dryas End Moraines towards the ice divide area in northern Finland. 1 = esker chain, 2 = Younger Dryas End Moraine. After Johansson et al. (2005).

THE ICE RETREAT IN THE FELL AREAS

The ice margin started to retreat from the Younger Dryas End Moraines about 11,600 years ago (Saarnisto 2000, Saarnisto & Saarinen 2001). At the same time the highest felltops in northwestern and northern Lapland appeared from the ice as nunataks. In the northern part of Lapland, the ice margin retreated towards south-southwest and in the southern part of Lapland to northwest. The surroundings of the Lake Inari were deglaciated approximately 10,800 years ago (Kujansuu et al. 1998) (Fig. 3).

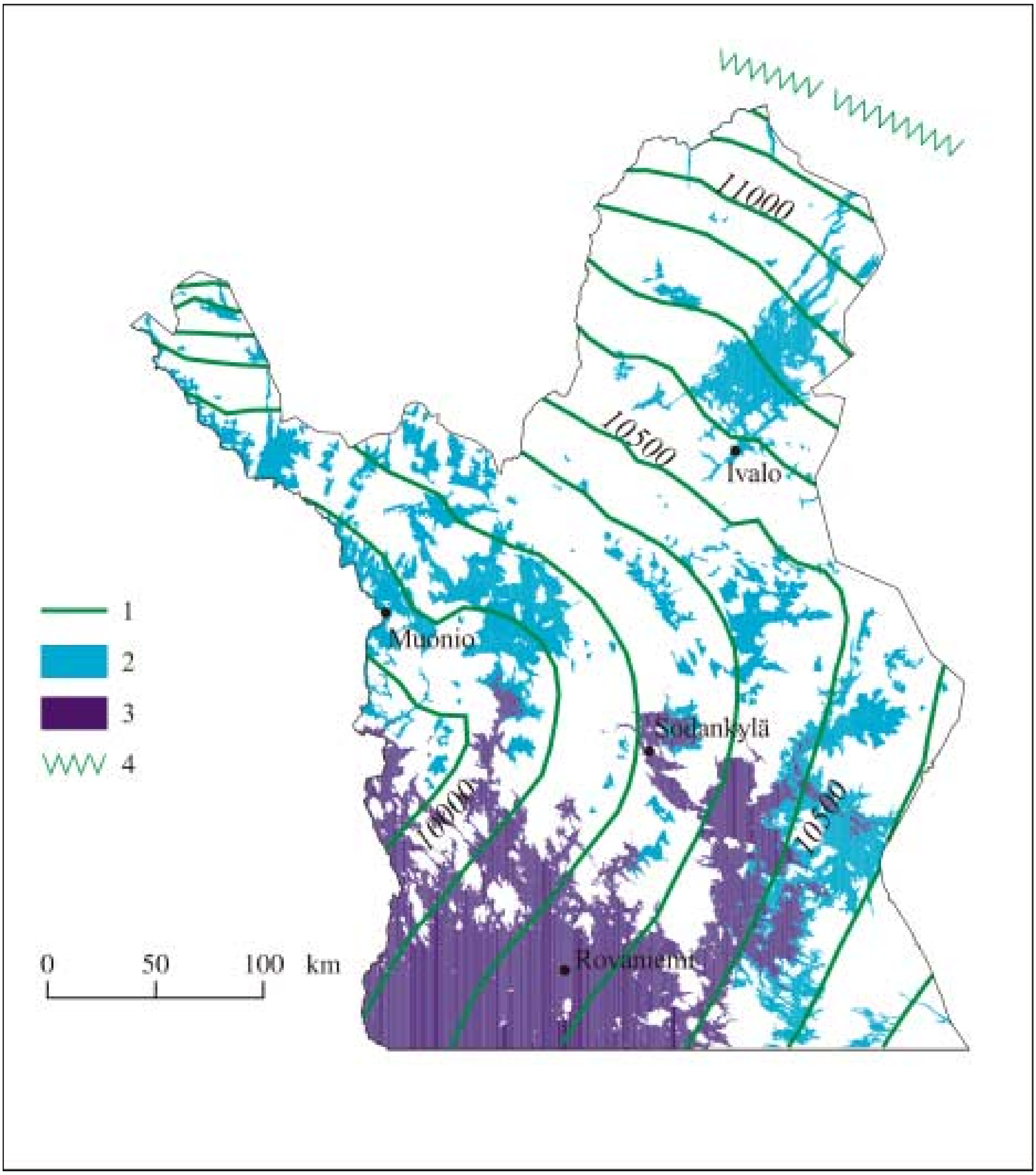

Fig. 3. Recession of the ice margin in northern Finland. 1 = position of the ice margin, 2 = areas covered by ice-dammed lakes, 3 = Ancylus Lake, 4 = Younger Dryas End Moraine. After Johansson et al. (2005).

Fig. 4. Lateral drainage channels on the slopes of Jäkäläpää Fell in Inari.

The ice margin reached the Lemmenjoki – Saariselkä fell range about 10,600 years ago. In the mountainous area the deglaciation was first thinning of the ice. During the nunatak phase many of the overflow channels in the Saariselkä area were also formed as a result of meltwater erosion. As the ice sheet became thinner and its surface sank, the nunataks expanded into vast ice-free areas. The ice flow stagnated behind the nunataks, and fractures and crevasses were consequently formed at the margin of the ice sheet. The meltwaters penetrated under the ice margin, eroded subglacial chutes and deposited engorged eskers (Johansson 1994).

The meltwater action at the boundary of the ice sheet and an exposed fell slope next to it produced series of shallow meltwater channels, i.e. lateral drainage channels (Fig. 4). The channels generally occur in groups, running side by side gently sloping down the fell side. These channels are of great importance in the study of deglaciation in minor scale, since they help to construct the position of the ice margin in a great detail, which gives a picture of the inclination of the surface of the ice sheet. Additionally they indicate that the ice margin was unbroken. In favourable places on the fell slopes channel systems comprising several tens of channels were formed in which the distance between the individual channels remains nearly constant. In these places the channels may have formed as a consequence of the annual thinning of the ice sheet and they can be used to document the rate of melting (Penttilä 1963, Johansson 1995). Their distribution indicates that the gradient of the ice surface was between 1.5 and 3:100 m and the ice margin thinned approximately 1.6 m per year. The rate of recession of the ice margin varied 130 – 170 m per year, approximately 140 m per year (Johansson 1995). In western Lapland the deglaciation rate was faster. Lateral drainage channels on the fellslopes in the Ylläs area show that the annual thinning was about 2.5 – 3 m and the ice retreat about 170 m per year (Abrahamsson 1974, Kujansuu 1967). The melting accelerated due to the decreasing mass of the continental ice sheet as the deglaciation proceeded and to warming of the climate, which also contributed to the acceleration.

As the retreat of the ice margin continued, the southern edge of the Lemmenjoki – Saariselkä fell range became exposed. The ice margin was pressing against the fell range and it became undulated with tens of kilometres long ice lobes penetrating into the valleys of Lemmenjoki, Suomujoki and Anterijoki. At the end of this ice-lobe stage, the lobes turned passive and melted down after they had lost contact with the active ice sheet.

ICE LAKES AND EXTRAMARGINAL CHANNEL SYSTEMS

To the south of the Lemmenjoki – Saariselkä fell range ice lakes were dammed in front of its margin during various stages of deglaciation. A requirement for the formation of an ice lake was that the ice margin retreated downslope along the main river valleys. The glacier had to be active and solid at its edges to be able to dam any meltwater. In such cases the meltwaters were not able to drain but formed ice-dammed lakes in the valleys. The different lake phases are indicated by the presence of raised shorelines and outlet channels, coarse outwash sediments and fine-grained glaciolacustrine sediments.

While the ice sheet was still leaning against the southern slope of the Lemmenjoki-Saariselkä fell range, several ice lakes started to form in depressions between the fell tops (Tanner 1915, Penttilä 1963, Johansson 1995). They were short-lived and small, but since they occupied valleys between fells they became deep, and they contained a considerable volume of water. On even terrains large and shallow ice lakes were formed. The largest of these ice lakes were located in Salla, eastern Lapland, in the Muoniojoki and Ounasjoki river valleys, western Lapland, and in the Inari Lake basin, northern Lapland (Fig. 3). They covered thousands of square kilometres. However, most of the ice lakes were small, occupying only the deepest parts of the river valleys. The ice lakes drained from one river valley to another across the water divides creating erosional landforms, i.e. narrow and deep overflow channels and broad and shallow marginal and extramarginal channels. Successive marginal and extramarginal meltwater systems formed along the retreating ice front and some of them can be followed hundreds of kilometres from the higher terrain in northwestern Lapland to the lower levels in eastern Lapland (Kujansuu 1967, Lunkka et al. 2004, Johansson & Kujansuu 2005). Initially the meltwater flowed northwards, across the main water divide towards the Arctic Ocean. As the ice sheet became smaller and retreated southwestwards the meltwater flow switched eastwards towards the White Sea and finally southeastwards along the retreating ice margin towards the Baltic Basin. The extensive marginal and extramarginal channel systems indicate that the ice sheet remained dynamically active until it wasted away.

In eastern Lapland the deglaciation was accompanied by the damming of the largest ice lake of northern Finland, the Salla Ice Lake, in the area between the present water divide and the ice margin (Johansson 1995). The water from the Salla Ice Lake flowed east through the Kutsajoki river to the White Sea. When the new outlet channel, which led to Korouoma, had opened under the margin of the retreating glacier about 10,500 years ago, the surface of the ice lake sank by more than 50 m in a short time. Extensive areas near the ice margin turned into dry land. This may have had a significant effect on the stagnation of the ice margins, the ice retreat and on the formation of the areas of hummocky moraines and ribbed moraines in the vicinity of Rovaniemi and Kemijärvi (Kurimo 1978, Taipale & Saarnisto 1991, Sarala 2005).

DISAPPEARANCE OF THE GLACIER IN THE ICE-DIVIDE AREA

When the margin of the retreating glacier had reached the ice-divide area in northern central Lapland about 10,300 years ago, it stagnated in several places at Posoaapa and Koitelainen areas and melted in situ, partially as separate patches of dead ice. Landforms typical of subglacial and marginal meltwater processes such as eskers and various channels are almost entirely missing. This shows that the marginal parts of the glaciers were broken and not capable of controlling the meltwater systems and the shallow ice lakes were obviously largely filled with ice blocks. In western part of the ice-divide area the ice margin was sufficiently intact to dam ice lakes. It retreated mainly to the west and at the final stage to the southwest. Some erosional remnant fells, which rose above the even terrain, such as Kumputunturi (Johansson et al. 1996), Aakenustunturi and the fell range of Pallas–Ounastunturi controlled the ice flow locally and divided the ice margin into separate lobes. West of Pallas- Ounastunturi the margin retreated to the south-southwest and on the east side to the south or south-southeast (Kujansuu 1967).

Glacial processes and morphology give a reliable picture of the ice recession but offer no basis for dating. The deglaciation of northern Finland lasted about 1,500 years and its beginning (11,600 years ago) can with certainty be connected with the end of the Younger Dryas chron. The last remains of the glacier disappeared from Kolari, western Finnish Lapland as the ice margin retreated to northern Sweden, which according to estimation occurred about 10,000 years ago (Saarnisto 2000, Lundqvist 2004) (Fig. 3). Radiocarbon dating of the bottom layers of peat deposits has been used to get an approximate age of the earliest vegetation, but these ages are inaccurate, mainly too old, to represent the final stage of deglaciation (Ignatius et al. 1980). After the retreat of the glacier the southern part of northern Finland was to a great extent covered by the Ancylus Lake. Due to land uplift the highest shore level formed at this phase is now at an altitude of over 200 m.

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LATE WEICHSELIAN SHEETFLOW DRAINAGE OF SUBGLACIAL LOKKA ICE LAKE

by
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Sutinen, R., Jakonen, M., Liwata, P. & Hyvönen, E. 2007. Late Weichselian sheetflow drainage of subglacial Lokka ice lake. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 55–62, 6 figures.

A conceptual model regards arboreal eskers as products of time-transgressive glaciofluvial deposition within close proximity of the retreating ice margin. In contrast, an anastomosing network of tunnel valleys or esker tracks is attributed to subglacial flood events, but the outburst release mechanisms of former subglacial meltwater bodies have remained obscure. We studied an anastomosing network of esker tracks in Lokka, east-central Finnish Lapland. Sheetflow paleohydrography as compiled with a digital elevation model (DEM) and airborne gamma-ray (AGR) data disputes the concept of ice-marginal deposition, but the network displays evidence of subglacial mass-flows. The tracks show convex-up parts in their long profiles and cross topography regardless of slope indicating a subglacial origin beneath the Fennoscandian Ice Sheet (FIS). The alignment of the esker tracks, fanning out to southeast and south seems to be generated from a large subglacial reservoir beneath the FIS, and we propose that the former Lokka ice-lake had been the source of the sheetflow system. The coincidence of escarpments (DEM) and airborne magnetic (AM) anomalies suggests that neotectonic fault instability likely was the triggering mechanism for the outburst of the subglacial Lake Lokka.

Key words (GeoRef Thesaurus, AGI): glacial geology, subglacial lakes, meltwater, drainage patterns, eskers, neotectonics, Pleistocene, Weichselian, Lokka, Lapland Province, Finland.

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INTRODUCTION

Eskers and tunnel valleys have been understood to be meltwater forms, but the organization of these landforms, however, exhibits evidence of different meltdown drainage patterns of Pleistocene ice sheets. The conceptual model regards eskers as products of time-transgressive glaciofluvial deposition within close proximity of the retreating ice margin (Banerjee & McDonald 1975). On a subcontinental scale, arboreal esker systems also show time-transgressive features (Clark & Walder 1994). Eskers are commonly associated with anastomosing tunnel valleys that are found to be common in regions typified by fine-grained tills derived from sedimentary bedrock (Wright 1973, Clark & Walder 1994). It is assumed that high-velocity flood event(s) preceded the evolution of eskers found in the tunnel valleys (Wright 1973, Kozlowski et al. 2005). Morphological evidence has also shown that catastrophic sheetflood events are capable of eroding large regions out of pervasive cover of unconsolidated sediments (Kor & Cowell 1998). The anastomosing network of tunnel valleys, esker tracks or sheetfloods are evidently attributed to subglacial lakes (Wright 1973, Shaw & Kvill 1984, Sutinen 1992, Kor & Cowell 1998, Sjogren et al. 2002, Kozlowski et al. 2005), but the outburst release mechanisms of former subglacial meltwater bodies have remained unknown.

Radar evidence has indicated large subglacial lakes (e.g. Lake Vostok, 500 by 300 km in size) beneath the Antarctic Ice Sheet (AIS) (Robin et al. 1970, Kapitsa et al. 1996, Siegert 2000), hence potential water volumes for tunnel valley formation evidently existed beneath the former Pleistocene ice sheets. Following the concept of glacio-seismotectonics at the margin of FIS (Stewart et al. 2000), we suspect that large-magnitude seismic impacts (Arvidsson 1996, Wu et al. 1999) have been part of an abrupt release of subglacial water bodies as outburst sheetflows beneath the FIS. In this study, we present morphological/geophysical evidence that neotectonic fault instability likely contributed not only to paleolandslides (Sutinen 1992, 2005, Sutinen et al. this issue), but also to glaciofluvial landform genesis shortly after the Younger Dryas (YD) episode in Finnish Lapland.

MATERIALS AND METHODS

We compiled DEM and low altitude AGR (γ_K) to outline the anastomosing network of sheetflow tracks within an area of 100 by 100 km in east-central Finnish Lapland (Fig. 1). The time-stability of spatial structures in the soil γ_K (Hyvönen & Sutinen 1998) implies that the terrain anisotropy, whether due to glacial streamlining or sheetflow outbursts, is revealed by the AGR-γ_K. The regimes of low soil volumetric water content (θ_v), such as sheetflow esker ridges, are indicated by high γ_K-signatures, while low γ_K-signatures (high θ_v) typify topographic lows, often with organic or fine-grained materials (Hyvönen et al. 2003, Sutinen et al. 2005). The DEM-AGR enhancement of the sheetflow associations was viewed closer in the subset areas of Tanhua and Suoltijoki (Fig. 1.).

Compilation of AM and DEM was applied to search candidate lines to represent neotectonic faulting. The coincidental appearance of topographic escarpments and AM-anomalies, showing, for example, pre-existing hydrothermal alteration zones, marks potential zones where faults were likely reactivated within the glacio-seismotectonic subsidence-rebound processes (Arvidsson 1996, Fjeldskaar et al. 2000, Stewart et al. 2002). However, not all escarpments are neotectonic and not every AM-anomaly is present at sites with neotectonic faults (Ojala et al. 2004). Thus, glacio-morphological investigations are often needed to verify neotectonic origin (Lagerbäck 1990, Dehls et al. 2000). For the time-stratigraphic purposes, additional data on the preferred orientation of till(s) were acquired by applying non-destructive azimuthal measurements of apparent electrical conductivity (σ_a) (Taylor & Fleming 1988, Penttinen et al. 1999, Sutinen et al. this issue). We determined the maximum σ_a-anisotropy (orientation) of the tills at 12 sites with a Geonics EM38 (Geonics Ltd. Mississauga, Ont. Canada) device designed to a depth range of 1.5-2 m (depending on material) in vertical coil position EM38(V). The field campaign was carried out in August 2005.

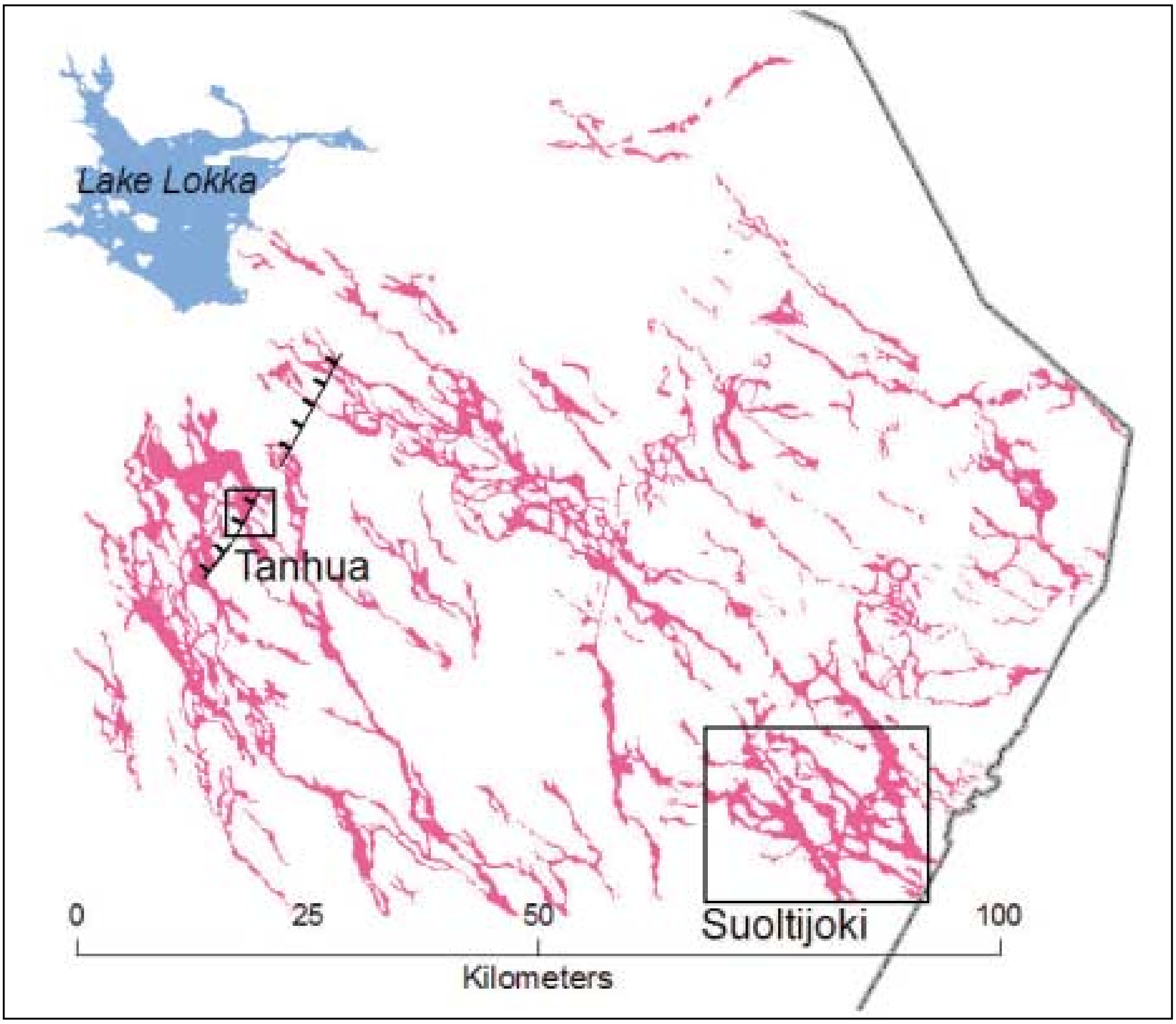

Fig. 1. Anastomosing network of sheetflow tracks in the Lokka area, east-central Finnish Lapland. The Tanhua and Suoltijoki subsets are viewed in more detail in the text. The Tanhua fault line (tentatively Late Glacial Fault; LGF) shown with barbs turned towards the lower block. Descriptions of pre-Late Weichselian NE-W (top of the figure) and N-S (in the mid of the figure) oriented esker tracts were given in Sutinen (1992).

RESULTS AND DISCUSSION

Time-stratigraphic context

The alignment of sheetflow tracks, fanning out to the southeast and south indicates drainage of large subglacial reservoir beneath the FIS after YD (Fig. 1). The former Lokka ice-lake most likely was the source for the sheetflow system. The sheetflow pattern crosses the N-S oriented esker system (Fig. 1), previously interpreted as Mid-Weichselian by Sutinen (1992). Also, maximum σ_a-anisotropy of the tills exhibited a rather concurrent pattern with the N-S eskers (Fig. 2) suggesting that the N-S oriented eskers were generated, probably through time-transgressive evolution (Banerjee & McDonald 1975), prior to the sheetflow drainage of Lake Lokka. The preservation of Mid-Weichselian eskers and till-sequence indicates that the erosion rate of the Late Weichselian ice-flow had been low in the Lokka region. Morphologically this is supported by the lack of erosional streamlined features commonly found inside the YD End Moraines (YDEMs; Nordkalott Project 1987). We also argue against high velocity flood events during the sheetflow drainage because large regions had not eroded out of unconsolidated sediments (cf. Kor & Cowell 1998).

Sheetflows vs. arborescent eskers

The anastomosing pattern of sheetflow tracks in the Lokka region (Figs. 1 and 3) is different from those in lower parts of northern Fennoscandia and the Canadian Shield displaying river-like drainage pattern of widely spaced tributary branches. The arborescent esker chains can often be followed tens or even hundreds of kilometres (Nordkalott Project 1987; Clark & Walder 1994). According to Clark and Walder (1994), anastomosing or braided channel systems are attributed to sedimentary terrains with fine-grained and low-permeability tills, not high-permeable tills underlain by crystalline bedrock as is the case in Lokka region. We found network patterns of some eskers (Fig. 4) rather similar to those described earlier by Prest (1968) and, Shaw and Kvill (1984) in Saskatschewan, Canada. However, discontinuities or even offsets in the esker tracks are difficult to explain without subglacial fault instability.

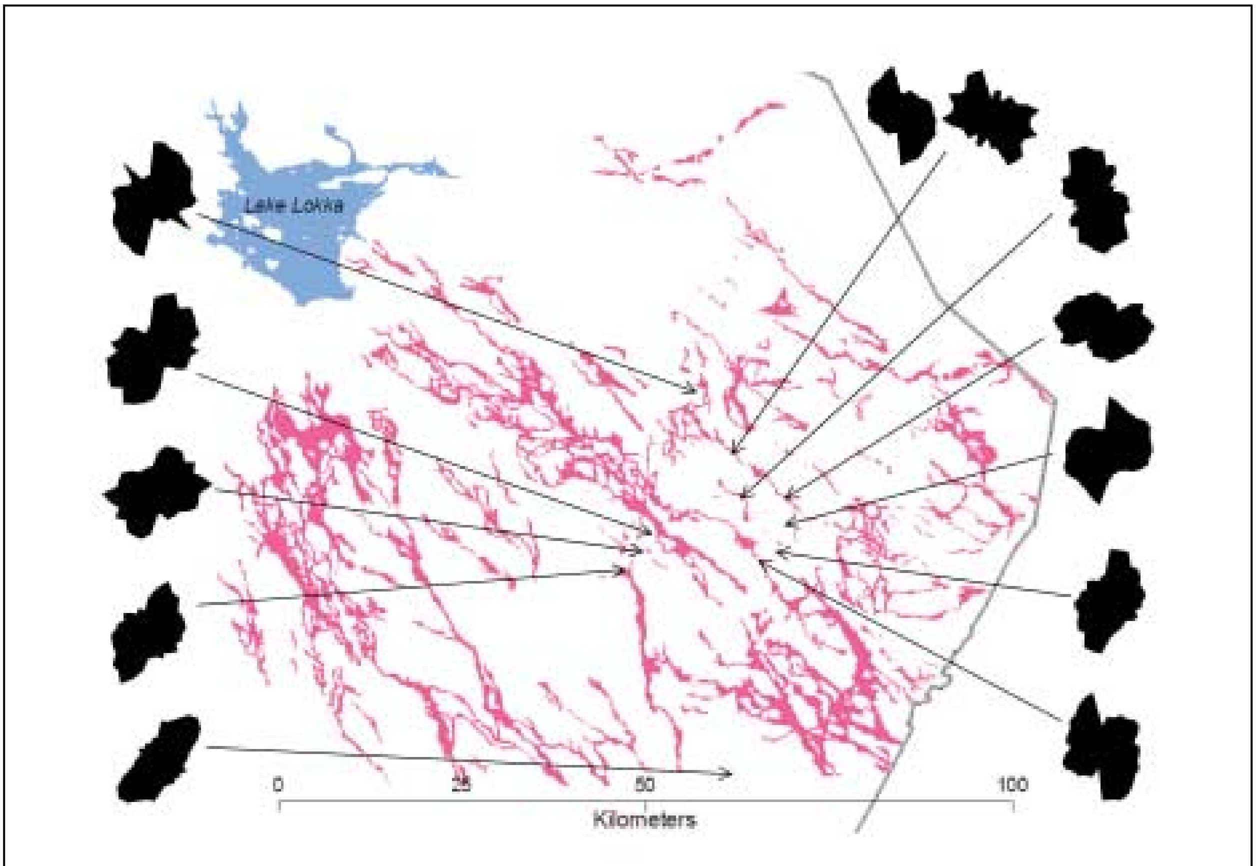

Fig. 2. Anastomosing network of sheetflow tracks crosses the N-S-oriented esker ridges in the Lokka area. Note the majority of maximum σ_a-anisotropy of tills oriented transversally to the sheetflow system, but rather than concurrently with the N-S eskers.

In Lokka, the sheetflow tracks were found to be less than 20 km wide and less than 60 km long, but were minor in cross-sectional channel form (cf. Wright 1973). The tracks show convex-up parts in their long profiles and cross topography regardless of slope. The tracks may cross the topography with 8% of slope displaying evidence of subglacial origin. The tracks contained both erosional channels and esker ridges. In some cases, the tracks exhibited 3-12 parallel or subparallel ridges 5-10 m in height

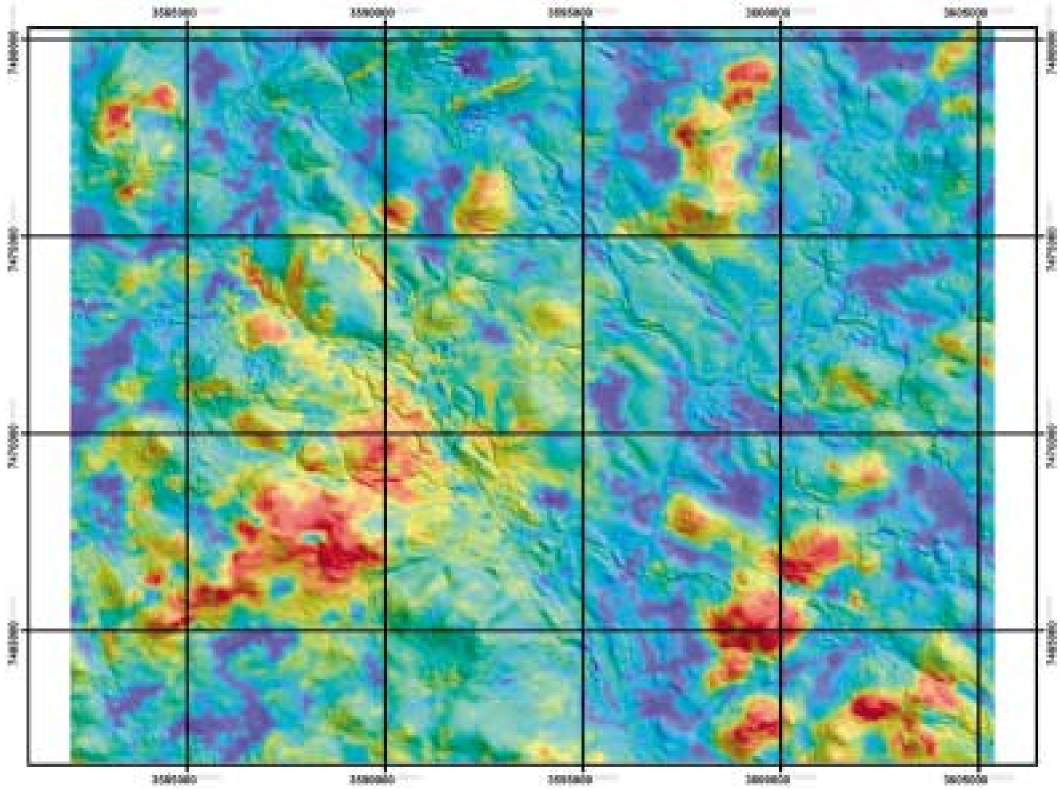

Fig. 3. A compilation of DEM (data from National Land Survey) and low altitude AGR (γ_K; by Geological Survey of Finland) outlining the anastomosing network of sheetflow tracks within the Suoltijoki subset area. The grid size is 5 by 5 km.

Fig. 4. A downflow outlook of the network of parallel/ subparallel esker ridges in the Suoltijoki subset area.

(Fig. 4). Hence, the drainage mechanism is dissimilar to subglacial tunnel valleys as having wide and steep walled profiles (Wright 1973, Kozlowski et al. 2005) and sometimes containing hummocky moraines on the bottom of the channels (Sjogren et al. 2002). The original concept of tunnel valleys in Minnesota by Wright (1973) suggests that the meltwater reservoir likely existed in Hudson Bay and the valleys had been carved by high-velocity subglacial streams driven by the hydrostatic pressure under active ice. However, as the ice thinned to stagnation the hydrostatic head was lost and small eskers along the tunnel valleys were deposited (Wright 1973). This was not the case with the sheetflow tracks in Lokka because the formation of a braided system of tunnel valleys is unlikely in a region typified by thin drift underlain by crystalline bedrock (Clark & Walder 1994). Even though the outbursts and mass-flows of debris appeared to be due to abrupt events, no high-velocity events had preceded the esker deposition, otherwise large regions had eroded out of unconsolidated sediments (Kor & Cowell 1998).

Mode of sheetflow deposition

The esker ridges within the sheetflow tracks (Figs. 1, 3-4) did not display evidence of time-transgressive evolution within the close proximity of retreating ice margin (Banerjee & McDonald 1975), but were generated by short-lived subglacial mass-flows. Hence, the mode of formation was different from those in openwork sedimentation (Banerjee & MacDonald 1975). The conduit infill deposits, found in the cores of eskers and involving flow of slurry material and heterogeneous suspension mode of full-pipe transport (Sutinen 1992) more likely was the mode of deposition in the sheetflow esker ridges in Lokka. Similarly to transition in particle size distribution between tills, mass-flow sediments (flow tills) and esker sediments in conduit infills (Sutinen 1992, see also Fromm 1965), we argue that transition from mass-flows to full-pipe flows occurred in the conduits. Comparable to cases in the Lokka region, the lack of openwork sediments on the surfaces of the esker ridges, subangular stones and blocks on the eskers and poorly sorted material are regarded as evidence of short transport distance of esker materials by Tanner (1915) in the Utsjoki region, Finland and by Fromm (1965) in Norrbotten, Sweden.

Neotectonic instability and Lake Lokka

The alignment of the network of sheetflow tracks suggests that a subglacial water body had most likely been present in the Lake Lokka basin (Fig. 1). The presence of lakes beneath the Pleistocene ice sheets is supported by radar evidence showing large subglacial lakes beneath the AIS (Robin et al. 1970, Kapitsa et al. 1996, Siegert 2000). However, the permanence of the Antarctic lakes seems to be synchronized with the seismic stability due to the equilibrium between the crustal subsidence and the overlying masses of the AIS (Johnson 1987). In contrast to the aseismic interior parts of the modern ice sheets (Johnson 1987), glacio-isostatic rebound made the crust highly instable at the retreating margin of FIS within and after YDEM (Arvidsson 1996, Wu et al. 1999, Fjeldskaar et al. 2000, Stewart et al. 2000). On the basis of the crustal subsidence-rebound model (Stewart et al. 2000), we suspect that fault instability was the driving force for the sheetflow outbursts of subglacial Lake Lokka beneath the FIS shortly after the YDEM-phases. One of the major fault lines (tentatively late glacial fault; LGF) was revealed with a compilation of AM and DEM (Fig. 5) in the Tanhua area (see location in Fig. 1). The vertical escarpment was clearly visible in the field with a lower block in the up-ice direction (Fig. 6), hence being directionally consistent with the postglacial rebound concept (Stewart et al. 2000). However, the appearance of the Tanhua escarpment is dissimilar to those within the PGFs, where previously deposited sediments (e.g. dating back to Younger Dryas) have undergone drastic earthquake deformations (Lagerbäck 1990, Dehls et al. 2000). Rather, the sheetflow tracks crossing the fault line, particularly without signs of collapsed esker forms, more likely were syndepositional with or shortly after the subglacial faulting.

The abundance of sheetflow tracks in the Lokka area (Fig. 1) has required the presence of a large amount of subglacial meltwater. The hydrology model by Arnold and Sharp (2002) suggests that the FIS in northern Fennoscandia was warm-based during deglaciation after 16,000 yr BP and a substantial amount of meltwater had discharged in 14,000-12,000 yr BP. Hence, potential volumes of subglacial water existed during the YDEMs. Subglacial lakes, capable of bringing about catastrophic outbursts, may also be due to volcanic eruptions (Björnsson 1998), but no evidence indicates Late Pleistocene volcanic activity in the Lokka region. It remains unclear if the sheetflows were due to one major fault deformation (see Fig. 1, 5-6) or if several faults transgressively contributed to subglacial meltwater outbursts (see Sutinen et al. this issue). Since the sheetflow tracks display discontinuities and sometimes offsets (see Fig. 1), multiple faults were apparently present and most likely concurrent with the maximum fault instability in Fennoscandia 11,000-9,000 yr BP (Wu et al. 1999).

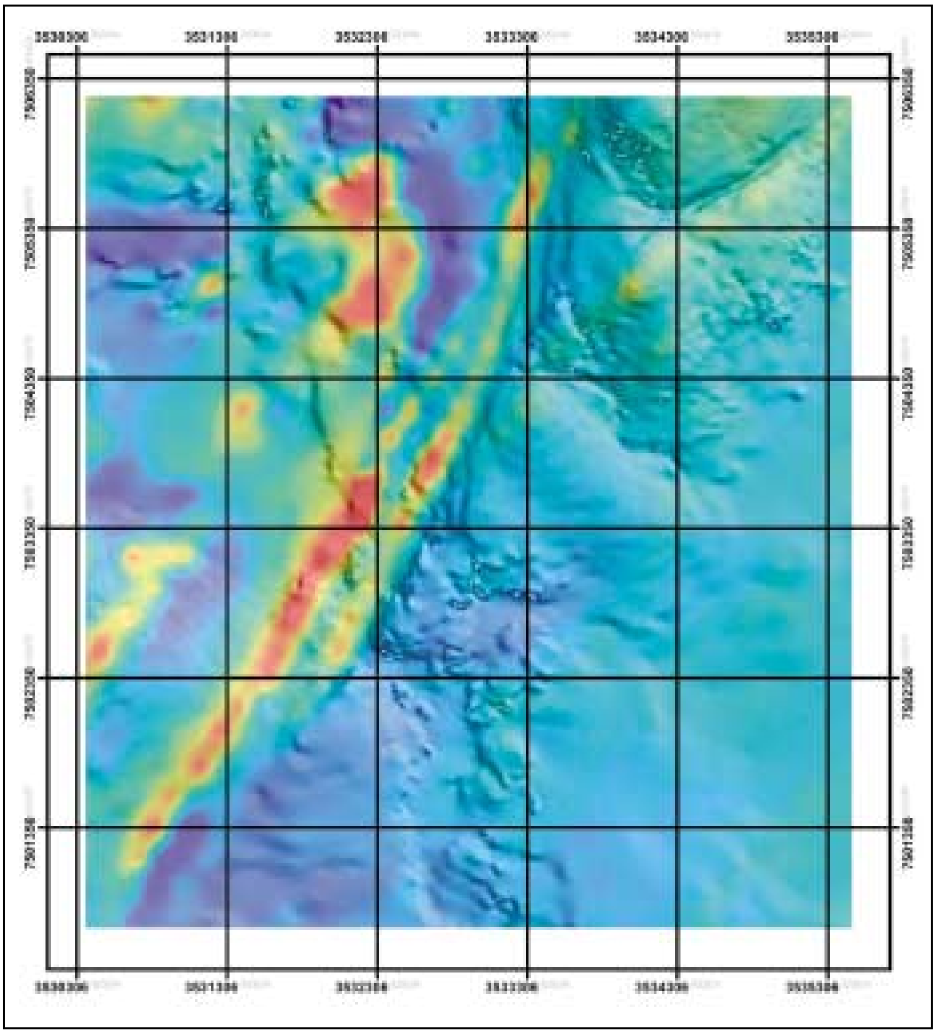

Fig. 5. A compilation of low altitude airborne magnetic (AM by the Geological Survey of Finland) and DEM (from National Land Survey) data coincidentally demonstrating fault lineament; here interpreted as Tanhua Late Glacial Fault (LGF). The grid size is one by one km.

Fig. 6. A view showing the Tanhua LGF escarpment (peat bog on the left towards the lower (up-ice) block. Note absence of sediments on the LGF escarpment and postglacial weathering of rock.

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SPATIAL DISTRIBUTION AND MORPHOLOGICAL ASPECTS OF ESKERS AND BEDROCK VALLEYS IN NORTH ESTONIA: IMPLICATIONS FOR THE RECONSTRUCTION OF A SUBGLACIAL DRAINAGE SYSTEM UNDER THE LATE WEICHSELIAN BALTIC ICE STREAM

by
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Rattas. M. 2007. Spatial distribution and morphological aspects of eskers and bedrock valleys in north Estonia: implications for the reconstruction of a subglacial drainage system under the Late Weichselian Baltic Ice Stream. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 63–68, 4 figures.

The spatial distribution of eskers and bedrock valleys, their antecedent role in draining subglacial meltwater and their relationship to bedrock lithology are discussed. Bedrock valleys and eskers are arranged in a sub-parallel pattern reflecting former meltwater flow towards the ice margin. Several esker systems are confined to the valley limits following either the valley floors or lying upon their shoulders. Such spatial relationships suggest that the initial focusing of meltwater may have been controlled by the location of bedrock valleys. Alternatively, some of the buried valleys or segments of valleys may be the result of tunnel valley formation. Some valleys have clear morphological characteristics of a tunnel valley, like abrupt termination and undulating or convex-up longitudinal profiles. Sedimentary records suggest that some valley infill and esker sediments may have been developed as a successive formation.

Key words (GeoRef Thesaurus, AGI): glacial geology, subglacial environment, meltwater, drainage patterns, eskers, valleys, Pleistocene, Weichselian, Estonia.

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INTRODUCTION

Subglacial meltwater storage and drainage have been identified as a key factor in determining rates of glacier flow and patterns and processes of subglacial sedimentation. Warmed-based Pleistocene ice sheets appear to be drained of their basal meltwater either by channelized flow or as groundwater flow (Benn & Evans 1998). The subglacial tunnel discharge and groundwater flow are coupled depending on water pressure gradients, primarily controlled by the basal melting rate and bed transmissivity (Boulton et al. 2001, 2003). The quantity of meltwater that can flow away as groundwater determines how much is left for tunnels, and it is therefore one of the key issues in understanding the spatial distribution of eskers and tunnel valleys.

Fig. 1. A general geological map of Estonia showing the distribution of bedrock valleys (by Suuroja 1997), eskers (by Kajak 1999) and the Late Weichselian ice margins (by Kalm 2006).

Estonia plays a significant role in the former meltwater drainage characteristics and pathways in the region of the Scandinavian Ice Sheet, between the Fennoscandian Shield and the Paleozoic sedimentary rock basin. Typical esker systems occur only in the northern and western parts of the territory associated with less permeable Ordovician and Silurian carbonaceous bedrock. Further to south, in areas of more permeable Devonian siliciclastic rocks (mostly sandstone) typical esker systems are less common and subglacial water has to be drained through the permeable bed or through the bedrock/tunnel valleys (Fig. 1). However, following the bedrock lithology, we see two totally different bedrock valley systems; in the carbonaceous bedrock area the valleys are distributed parallel to one another and arranged mostly from the northwest to southeast; in the Devonian sandstone area, they have a more uneven distribution forming a complicated jointed drainage network.

The broad aim of the ongoing project is to evaluate the meltwater drainage system beneath the Weichselian Baltic Ice Stream and to determine the role of subglacial tunnels in groundwater flow patterns during the last glaciation period in Estonia. Further details of studies include the identification and documentation of subglacial tunnels and their morpho-sedimentary relations to evaluate the development of the subglacial drainage system and to construct a relative chronology for the last deglaciation events in the region. The distribution and morphology of eskers and bedrock valleys has been thoroughly mapped (Kajak 1999, Suuroja 1997), but the nature of sediments and hydrologic characteristics of the tunnels are still poorly defined. Up to now, no direct interpolations have been made to identify which of the bedrock valleys, both buried and open, can be considered to be of subglacially meltwater-eroded origin. In this paper, the spatial distribution of eskers and bedrock valleys, their antecedent role in draining subglacial meltwater and their relationship to bedrock lithology will be discussed.

BEDROCK VALLEYS

In northern Estonia, bedrock valleys appear in the morphology of the incised substratum either at the present land surface or are buried under younger sediments (Fig. 2). The origin of the valleys has been a matter of controversy and in many cases, a composite origin should be considered. A number of valleys originated from the pre-Quaternary river drainage (Miidel & Raukas 1991), but were definitely heavily modified during the glaciation cycles. Older, pre-existing valleys have been re-used as subglacial drainage pathways and preferentially eroded over and over again during several glaciations. Individual valleys can be characterized as rectilinear to slightly sinuous, mostly orientated parallel or near parallel to the direction of ice movement, that is northwest to southeast. Thus, direct glacial erosion (quarrying) is considered to have played the most important role on their formation (Tavast & Raukas 1982). Occasionally the valleys are abrupt, with up to 90° angld bends that appear to be related to the bedrock fracture zones. Maximum valley depths reach about 150 m (to about 150 m a.s.l.) in the fore-clint area, but in the area of carbonate plateau generally not more than 10-40 m. The width is often 200-600 m, but may range up to 2 km in the fore-clint area. The cross-sections are mostly U-shaped with rather steep sides and flat floors. Often the valleys contain postglacial streams, lakes and small bogs along the valley floors (Fig. 2).

The sedimentary infill of deeper buried valleys consists of several till beds stratified with aqueoglacial deposits, and may be as old as Saalian, and in places, Eemian interglacial deposits (Raukas 1978, Kadastik et al. 2003). Shallow valleys often contain only meltwater deposits lying directly on the valley floor and with a relatively coarse grained material in the lower unit. The morphological characteristics of these valleys suggest strong erosion by pressurized water in the tunnels. Longitudinal valley profiles are slightly irregular with multiple lows and thresholds. Some valleys begin and then terminate rather abruptly.

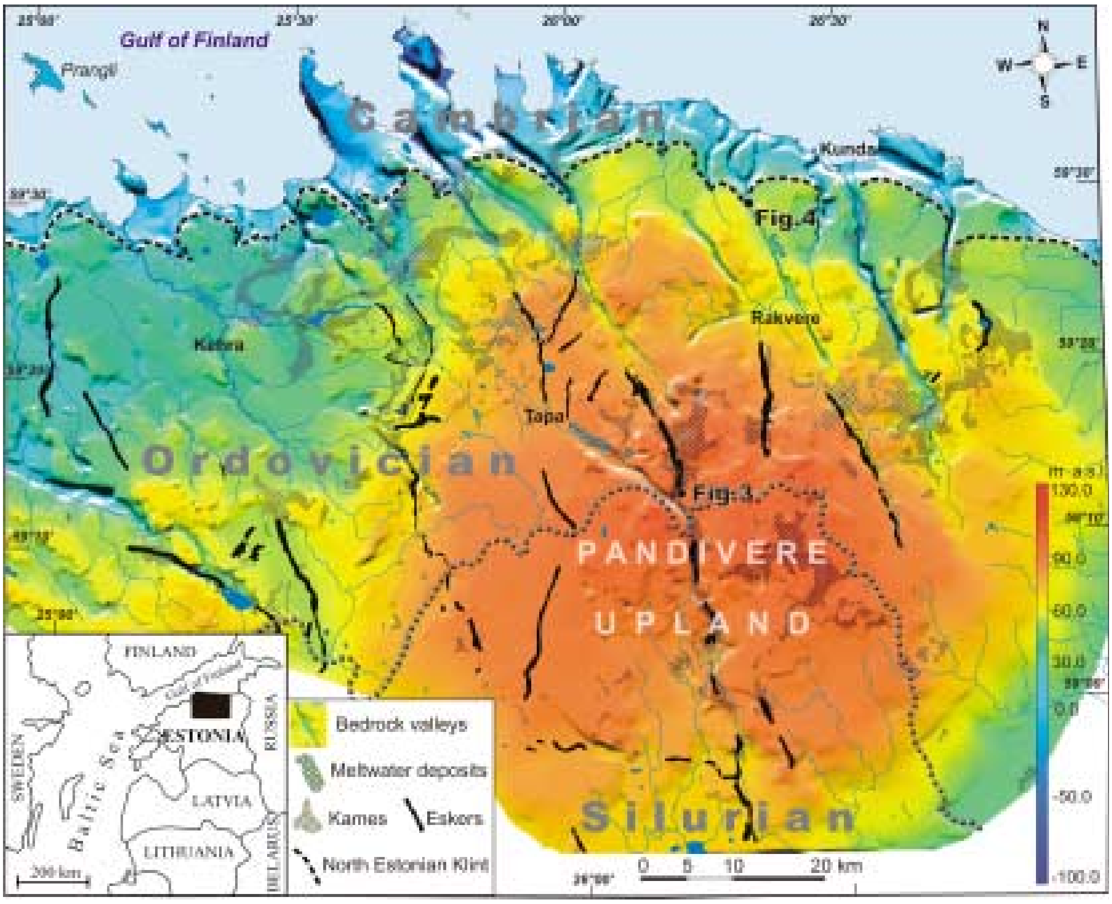

Fig. 2. Digital elevation model (DEM) of the bedrock surface showing the relationship between bedrock valleys and eskers in northern Estonia. DEM is constructed from geological mapping data with 5 m interval isolines. Postglacial streams and lakes often occur in bedrock valleys.

ESKERS

Eskers in Estonia are located in the northern part of the territory and are associated with Paleozoic carbonaceous bedrock. The morphology and sedimentology of eskers in Estonia were investigated in more detail in the 1960s (Rähni 1957, 1960, Raukas et al. 1971, cf. Karukäpp 2005). Entire esker ridges can vary from tens of metres to several kilometres in length. They are generally less than 500 m in width, but may exceed one kilometre. The relative height reaches up to 35 m, but is usually 10-20 m, and the gradient of the slopes is up to 40° (Raukas et al. 1971). The largest esker system crosses almost north to south and the central part of the Pandivere Upland has a length of about 60 km (Fig. 2). Eskers show a large variation in facies characteristics indicating non-uniform sedimentation due to fast changes in meltwater volume, energy and velocity.

Two general types of esker systems have been observed – those that have formed subglacially in tunnels and those that appear on the ice-marginals in deltaic or fan environments. Earlier investigations suggest that the majority of the radial eskers were formed in the open crevasses of the dead or passive ice (Raukas et al. 1971). There has been significant development in glacier hydrology and glacial sedimentology, and revision of the definition and classification of Estonian eskers is needed regarding the morpho-sedimentary relations of eskers and the hydraulic processes responsible for their formation.

DISCUSSION

A preliminary model of meltwater drainage pathways has been presented according to the spatial distribution of eskers and bedrock valleys (Rattas 2005, 2006). Several studies in the areas of the former ice sheets demonstrate that eskers and incised bedrock valleys are often interconnected (e.g. Attig et al. 1989, Fisher et al. 2005, Kozlowski et al. 2005, Jørgensen & Sandersen 2006). In northern Estonia, both bedrock valleys and eskers are arranged in a sub-parallel pattern, mostly from northwest to southeast, or sub-parallel to palaeo-ice streams reflecting former meltwater flow towards the ice margin (Fig. 2). Most of the esker systems are confined to the valley limits following either along the valley floors or lying upon valley shoulders or sometimes replacing a valley if it disappears over a short distance. Occasionally, some eskers cross valleys at rather sharp angles. This has been associated with late-glacial tectonic movements along tectonic fault zones (Rähni 1973) leading to the ice splitting and formation of open crevasses above the valleys. Considering that the main precondition for esker formation is low-permeability, and especially rigid substratum, then esker formation within valleys filled with permeable soft sediments, or above permeable tectonic fault zones could presumably be responsive to changes in subglacial hydraulic conditions leading to tunnel creation. When the tunnel forms, the water pressure rapidly decreases causing strong groundwater flow towards the tunnel. The large upward flow in tunnels and rapid pressure release seems to be responsible for carbonate precipitation. Such carbonate-cemented sand and gravel deposits are common in eskers and valley fills indicating their subglacial origin (Figs 3 and 4).

The sedimentary record suggests that some valley infill and esker sediments may have developed as a successive formation during periodical sedimentation by a number of meltwater flow episodes. Alternatively, some of the buried valleys or segments of valleys may be the result of tunnel valley formation with clear morphological characteristics of a tunnel valley, like abrupt termination and undulating or convex-up longitudinal profiles. This suggests that the initial focusing of meltwater may have been controlled by the location of bedrock valleys, and the steady-state discharges most likely preferred to follow pre-defined meltwater pathways.

Fig. 3. Poorly sorted cobble and boulder gravel are cemented by calcium carbonate due to groundwater discharge into the tunnel during the esker formation. See Fig. 2 for location (Photo: M. Rattas, 2002).

Fig. 4. Strongly cemented meltwater deposits on the slope of buried valley. See Fig. 2 for location (Photo: M. Rattas, 2002).

CONCLUSIONS

In northern Estonia, bedrock valleys and eskers are arranged in a sub-parallel pattern reflecting former meltwater flow towards the ice margin. The spatial relationships between bedrock valleys and eskers suggest that the initial meltwater focusing may have been controlled by the location of bedrock valleys. The steady-state discharges most likely preferred to follow pre-defined meltwater pathways. Sedimentary records suggest that some valley infill and esker sediments may have been developed as a successive formation. Alternatively, some of the buried valleys or segments of valleys may be the result of tunnel valley formation. Some valleys have clear morphological characteristics of a tunnel valley, like abrupt termination and undulating or convex-up longitudinal profiles.

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TUNNEL CHANNELS OF THE SAGINAW LOBE, MICHIGAN, USA

by
Alan E. Kehew[1] and Andrew L. Kozlowski[2]

Kehew, A. E. & Kozlowski, A. L. 2007. Tunnel Channels of the Saginaw Lobe, Michigan, USA. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 69–78, 10 figures.

During the last advance of the Laurentide Ice Sheet into southern Michigan, USA, the Saginaw Lobe advanced asynchronously with respect to the adjacent Lake Michigan and Huron-Erie Lobes, reaching its terminal position first, and then stagnating or retreating while the other lobes continued to advance.

A complex network of tunnel channels was formed in an intermittent and time-transgressive fashion beneath the Saginaw Lobe. One set of flow-parallel channels radiates from E-W to NE-SW to N-S orientations around the lobe. The channels typically exceed one kilometer in width and tens of meters in depth. In places, eskers occupy valley bottoms of the tunnel channels. Another set of channels cuts orthogonally across these channels at an approximate orientation of E-W to NW-SE.

Toward the southern limit of the Saginaw Lobe terrain, the channels were partially to nearly completely buried by ice and sediment deposited during collapse of the Saginaw Lobe and by sediment deposited during the overriding advances of the Lake Michigan and Huron-Erie Lobes. Relief in some channels subsequently increased as the buried ice melted. When the Lake Michigan Lobe began to retreat, the most recent set of tunnel channels, oriented trending westerly, were eroded by subglacial outbursts draining toward the former margin of the Saginaw Lobe.

Five genetic types of tunnel channels were identified in southern Michigan. These include: large, unburied channels (Type I), palimpsest channels partially buried by an end moraine formed by a readvance of the same lobe (Type II), palimpsest channels nearly completely buried by a readvance of the same lobe and not located near an end moraine (Type III), palimpsest channels partially buried by an advance of a different lobe than that under which the channels originally formed (Type IV), and unburied to partially buried channels that contain eskers (Type V).

Key words (GeoRef Thesaurus, AGI): glacial geology, glacial lobes, subglacial environment, meltwater channels, Pleistocene, Michigan, United States.

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INTRODUCTION

Recognition and descriptions of subglacially eroded channels formed beneath the former Laurentide and Scandinavian ice sheets have increased in recent years. These large, predominantly linear valleys are called tunnel channels when an origin by a single discharge of subglacial meltwater is inferred, or tunnel valleys when considered to form gradually or episodically, or when no genetic mechanism is implied (Clayton et al. 1999). Several defining characteristics of tunnel channels include: 1) dimensions reaching > 100 km in length, 4 km in width, and >3 m in depth, 2) generally straight reaches oriented parallel to the subglacial hydraulic gradient of the ice lobe, 3) undulating and adverse longitudinal gradients, and 4) irregular side slopes and valley bottoms (Kehew et al. 2007). The valleys appear to be most common near former ice margins, where they terminate at large glaciofluvial fans. They may be prominent topographic features or buried either partially or completely by younger glacial drift. Eskers may or may not be present on the valley bottoms.

Tunnel channels formed beneath the Scandinavian ice sheet have been described in German by Ehlers (1981), Grube (1983), and Piotrowski (1994). Jørgensen and Sanderson (2006) review the literature on Danish tunnel valleys. Most of the known tunnel channels associated with the Laurentide Ice Sheet are located in the glacial lobes of the Great Lakes region of North America. Wright (1973) first interpreted valleys in Minnesota as erosional forms of catastrophic subglacial discharges. Other studies have focused on tunnel channels in Minnesota (Mooers 1989, Patterson 1994, Hooke & Jennings 2006); Wisconsin (Attig et al. 1989, Clayton et al. 1999, Johnson 1999, Cutler et al. 2002); Michigan (Kehew et al. 1999, 2005, Fisher & Taylor 2002, Sjogren et al. 2002, Fisher et al. 2005, Kozlowski et al. 2005); and Ontario (Brennand & Shaw 1994).

Ó Cofaigh (1996) grouped the hypotheses for formation of tunnel channels into three main categories: (1) formation by subglacial sediment deformation, (2) gradual or time transgressive formation near ice margins, and (3) catastrophic subglacial sheet floods. The sediment deformation hypothesis is based on the work of Boulton and Hindmarsh (1987), who proposed that shallow channels carrying subglacial meltwater are initiated by piping from the ice margin and would gradually be enlarged by the creep of deformable sediment toward the channel and the subsequent removal of the sediment by meltwater flow. Opposition to this hypothesis is based upon arguments questioning the existence of a fluid-pressure gradient toward the channels and the occurrence of tunnel channels in bedrock lithologies, in which sediment deformation would not be possible (Ó Cofaigh 1996).

Time transgressive formation near ice margins is a hypothesis advocated by many authors. Formation was suggested by Mooers (1989) to be a more gradual, steady process, in which subglacial meltwater was augmented by diversion of supraglacial meltwater to the base of the glacier. On the other hand, most hypotheses invoke a sudden or catastrophic release of channelized subglacial meltwater because of the large clast size deposited in ice-marginal fans located at the termination of the channels (Piotrowski 1994, Cutler et al. 2002) and the size and dimensions of the channels (Kozlowski et al. 2005). The sudden release of subglacial reservoirs is often attributed to failure of a permafrost seal at the margin (Piotrowski 1994, 1997, Cutler et al. 2002, Jørgensen & Sanderson 2006, Hooke & Jennings, 2006). Piotrowski (1997) and Hooke & Jennings (2006) suggest a cyclical process in which a seal is punctured and leads to a catastrophic release of an impoundment, followed by the reestablishment of the seal and the refilling of the impoundment. Hooke & Jennings (2006) propose that piping and headward erosion back to meltwater impoundment initiate the outburst.

The catastrophic subglacial meltwater hypothesis for tunnel channels is one component of a complicated hypothesis for the origin of drumlins, Rogen moraine, megaflutes and other glacial landforms by broad subglacial sheetfloods (Shaw 2002). To account for the formation of landforms such as drumlins during the same megaflood that produced tunnel channels, advocates suggest the transition from sheetflow to channelized flow as the event progresses (Fisher et al. 2005). The sheetflood hypothesis is controversial, however, and has been challenged on various grounds (Clarke et al. 2005, Evans et al. 2006).

The presence of eskers in tunnel channels suggests that a depositional phase follows the erosional phase of channel formation, implying that the subglacial channel persists for a relatively long period of time. During deglaciation, collapse of ice and sediment into the channel may fill the channel for some period of time (Kehew et al. 1999, 2005, Kozlowski et al. 2005, Jørgensen & Sanderson 2006). Sediment from more recent glacial events may be deposited over the filled channel, leading to cross-cutting relationships that would be hard to explain without the presence of buried ice (Kehew et al. 1999). Gradual melting of the buried ice creates the final valley form, destroying the sharp, channel-like characteristics of the original valley.

GLACIAL LOBES IN MICHIGAN

The topography of southern Michigan, also known as the Lower Peninsula, is strongly controlled by the three lobes of the Laurentide Ice Sheet that occupied the area during the last (Wisconsin) glaciation. Ice advanced into Michigan around 26,000 ^{14}C yr BP and retreated for the last time around 11,000 ^{14}C yr BP (Larson & Schaetzl 2001). From west to east, the lobes include the Lake Michigan, the Saginaw, and the Huron-Erie Lobes (Figure 1). Both the Lake Michigan and Huron-Erie Lobes eroded deep troughs, now occupied by Lakes Michigan, Huron, and Erie. The Saginaw Lobe flowed southwesterly across the interior of the Peninsula, eroding Saginaw Bay of Lake Huron, which only extends partially along the flow path. Kehew et al. (2005) suggested that the Lake Michigan and Huron-Erie Lobes developed rapid, streaming flow, accounting for the size and depth of the troughs they created. By contrast, the central Saginaw Lobe advanced earlier into the area, but weakened or stagnated, as the flanking lobes grew stronger and encroached upon the recently deglaciated Saginaw Lobe terrain (Kehew et al. 1999, 2005).

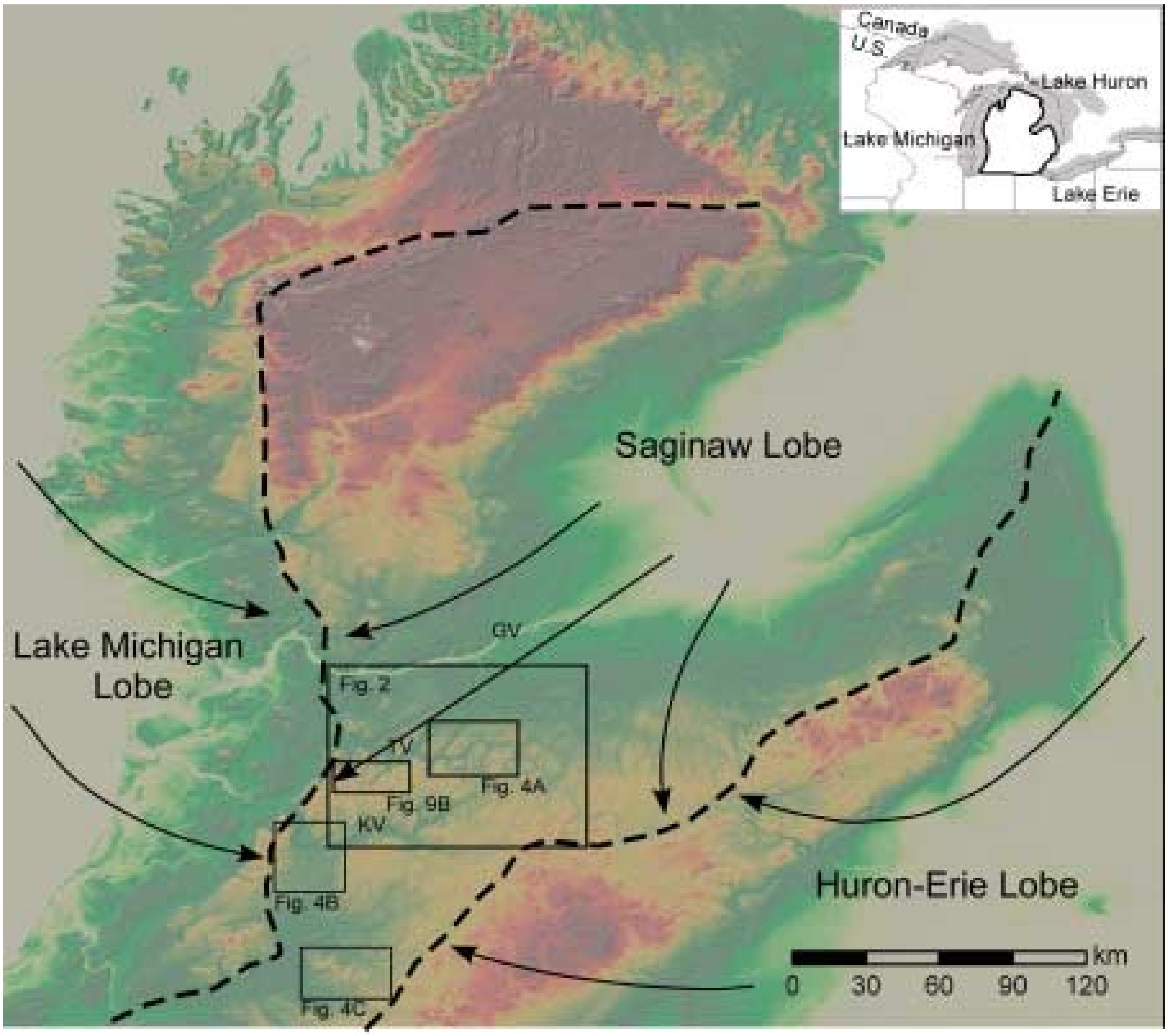


Fig. 1. Hillshade DEM of the Lower Peninsula of Michigan showing approximate extent of Lake Michigan, Saginaw, and Huron-Erie Lobes of the Laurentide Ice Sheet. Boxes show locations of other figures. GV: Grand Valley. TV: Thornapple Valley. KV: Kalamazoo Valley.

SAGINAW LOBE TUNNEL CHANNELS

The most distinctive feature of the Saginaw Lobe terrain is the network of sub-parallel channels that radiate outward from the center of the lobe. These channels are best developed down stream from the Saginaw Bay lowland, where they range in orientation from north-south in eastern part of the area, to northeast-southwest in the western part of the area (Fig. 2). These channels are interpreted as tunnel channels because of their orientation (parallel to the subglacial hydraulic gradient of the lobe), their undulating to adverse longitudinal profiles (Kozlowski et al. 2005) and the presence of eskers in some of the channels. The network of channels is cross cut by channels oriented nearly perpendicular to them (east-west), the largest of which are (from north to south), the Grand Valley, the Thornapple Valley, and the Kalamazoo Valley (Fig. 1). These channels are deeper and are interpreted as younger tunnel channels, although the Grand Valley was utilized as a proglacial lake spillway and enlarged from its presumed original size (Kehew 1993). The east-west reach of the Kalamazoo Valley, which was probably eroded subaerially, appears to have been fed by an anastomosing network of southwest-trending tributary tunnel valleys (Fig. 3) that conveyed meltwater to an ice margin. (Kozlowski et al. 2005).

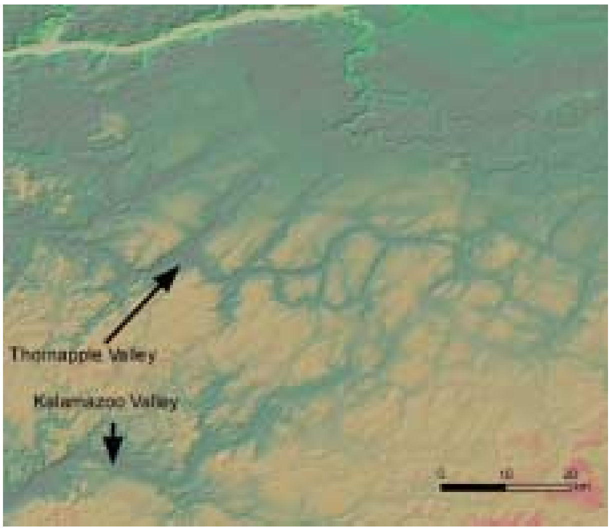


Fig. 2. Closeup view of tunnel channel networks of the central Saginaw Lobe.

South of the Kalamazoo Valley, the southwest-trending tunnel channels are much less distinct than the channels north of the valley. A major reason for this is that the channels appear to have been initially eroded when the Saginaw Lobe was at or near its Late Glacial Maximum (LGM) position, and then partially buried by sediment from later advances of the Saginaw Lobe or by sediment from the Lake

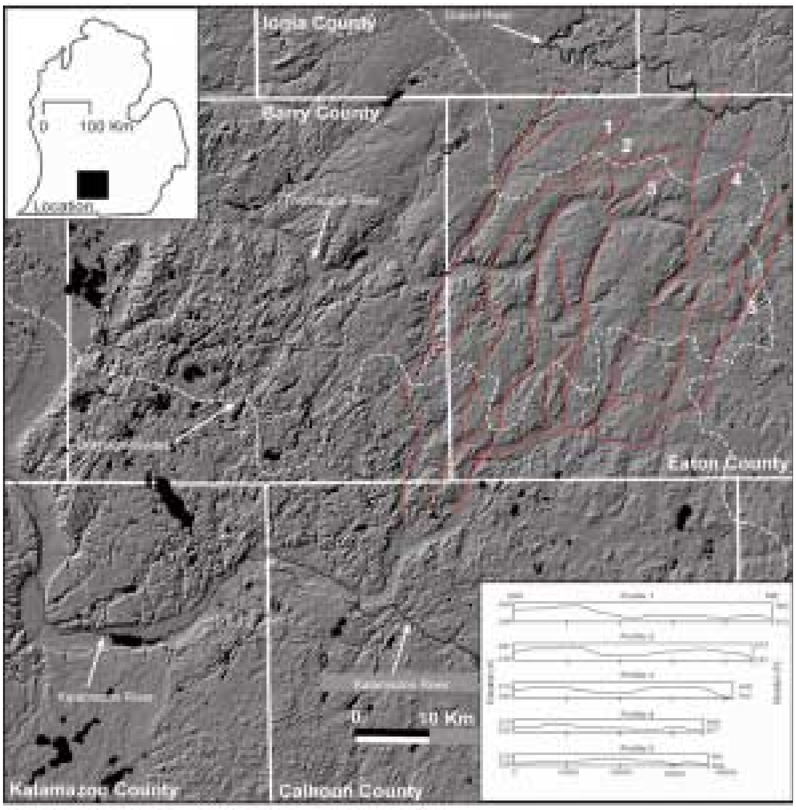


Fig. 3. Hillshade DEM displaying network of channels coalescing into the head of the central Kalamazoo Valley. Channels (black lines) interpreted as Saginaw Lobe tunnel channels. Dashed white lines are present day drainage divides. (From Kehew et al. 2007).

Michigan Lobe as it advanced into terrain vacated by the Saginaw Lobe. Because of their modification by subsequent advances, these tunnel channels have been described as palimpsest tunnel channels (Kehew et al. 1999). Cross-cutting relationships to be discussed below indicate that after formation, ice collapsed into the tunnel channel and became buried by debris, similar to tunnel channels in Denmark (Jørgensen & Sanderson 2006). This burial probably delayed melt out of the collapsed ice and formation of the valley for a period of hundreds to thousands of years.

EXAMPLES OF TUNNEL CHANNEL MORPHOLOGY AND INFERRED FORMATION

In this section, five examples of tunnel channels will be presented to illustrate the range in topography and post-erosional modification.

Type I Tunnel channels

Type I tunnel channels are generally large and display the least degree of burial (Figs. 4 and 5). They are interpreted to form by a subglacial burst of drainage to an ice margin followed by subsequent deglaciation without extensive collapse and filling with sediment. As a result, they are very distinct topographic features. Type I tunnel channels in the study area include both southwest and east-west orientations. The east-west channels cross cut the southwest trending channels and are interpreted to be younger.

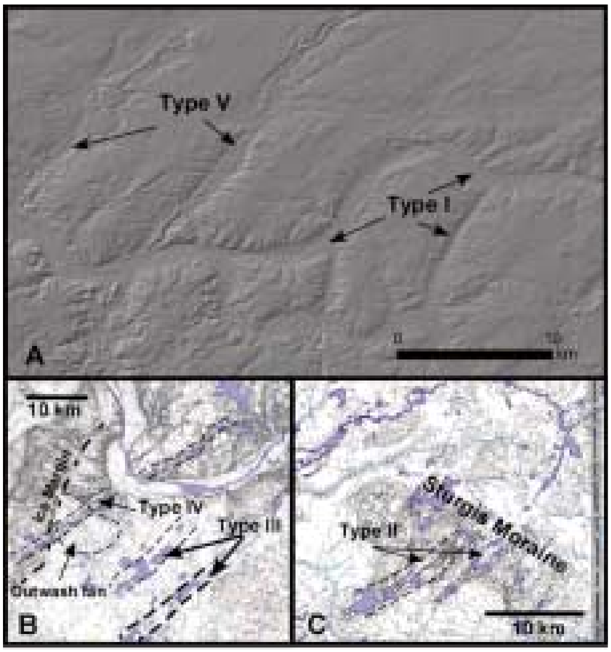


Fig. 4. Locations of types of tunnel channels discussed in text. Locations shown on Fig. 1. A. Hillshade DEM of Thornapple Valley. B. Contour map made from DEM (countour interval approximately 3 m.) Kalamazoo Valley is large valley in upper part of diagram. Ice margin is the Kalamazoo Moraine of the Lake Michigan Lobe. C. Contour map made from DEM.

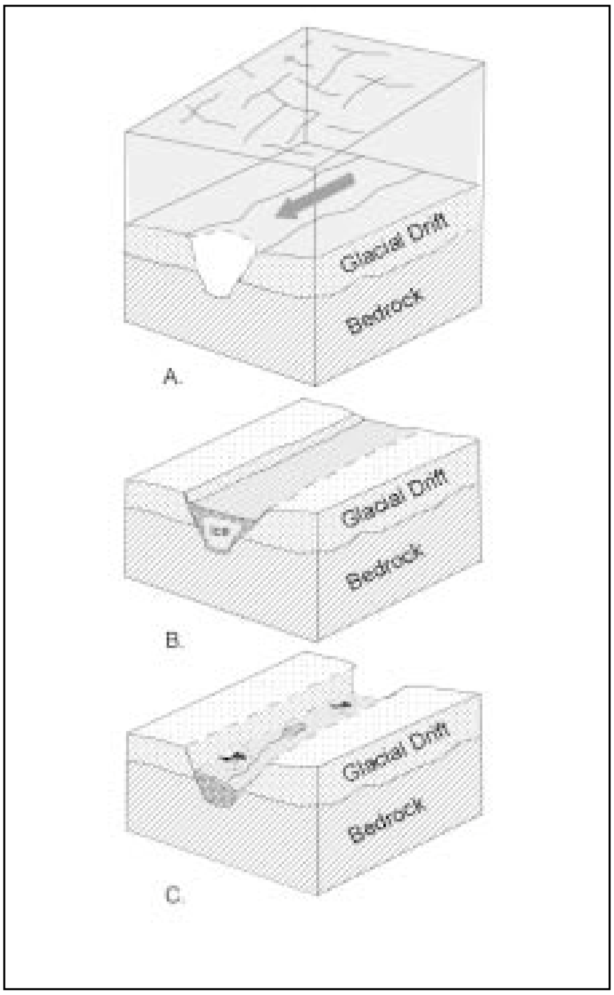


Fig. 5. Origin of Type I tunnel channels. A. Subglacial erosion of tunnel channel into bedrock. B. Remnants of collapsed ice and debris in channel after deglaciation. The amount of ice is relatively small in this case. C. Valley after meltout of buried ice. Valley walls are relatively straight and smooth.

Type II Tunnel channels

Type II tunnel channels (Figs. 4C and 6) are palimpsest channels that cut through terminal moraines of a more recent advance than the advance in which the channels were formed. The Sturgis Moraine of the Saginaw Lobe (Figs. 1 and 4C) is a good example of this type of tunnel channel. After erosion of the subglacial channel, it was likely filled to the top with collapsed ice and debris. Collapsed ice in the channel was still present at the time of the subsequent advance of the Saginaw Lobe to the position at which it formed the Sturgis Moraine. Melting of the buried, stagnant ice after formation of the moraine resulted in the development of indistinct valleys cutting through the moraine.

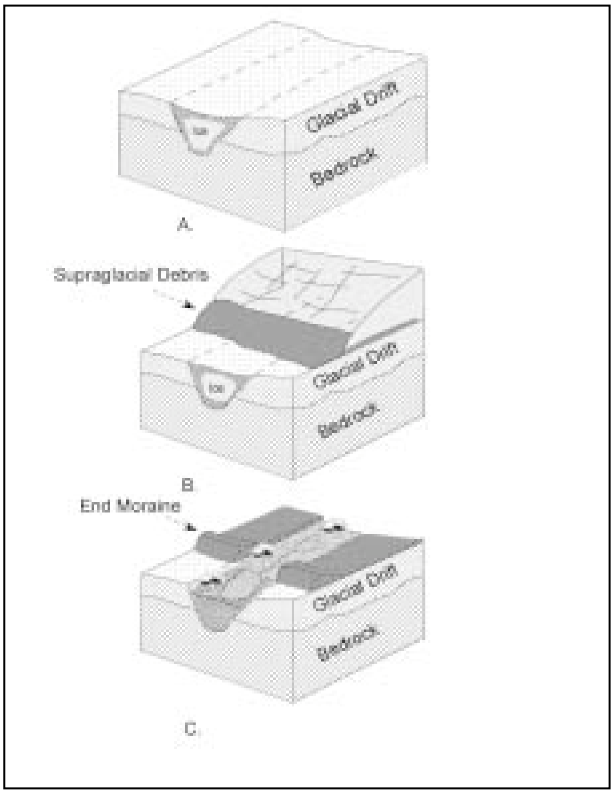


Fig. 6. Origin of Type II tunnel channels. A. Channel eroded as in Fig. 5A by early advance of Saginaw Lobe, filled with thick buried ice and debris, and deglaciated by retreat of lobe. B. Readvance of the Saginaw Lobe to form Sturgis Moraine at terminal position of readvance (Figs. 4C and 1). C. Gradual meltout and collapse of buried ice to form present-day valley, which cuts through the moraine.

Type III Tunnel channels

The Type III tunnel channel (Fig. 4B and 7) represents another type of palimpsest tunnel channel. Here, after filling of the valley by collapsed ice, a subsequent advance of the Saginaw Lobe completely overrode the channel, probably the same advance that formed the Sturgis Moraine. A more complex fill, involving a diamicton from the later advance, and surficial outwash, comprises the stratigraphy of the channel. Because of the degree of post-erosional filling of the channel, channels of this type are very subtle features on the modern landscape. The two Type III tunnel channels shown on Fig. 4B are recognizable only by an alignment of lake basins and no valley exits at land surface.

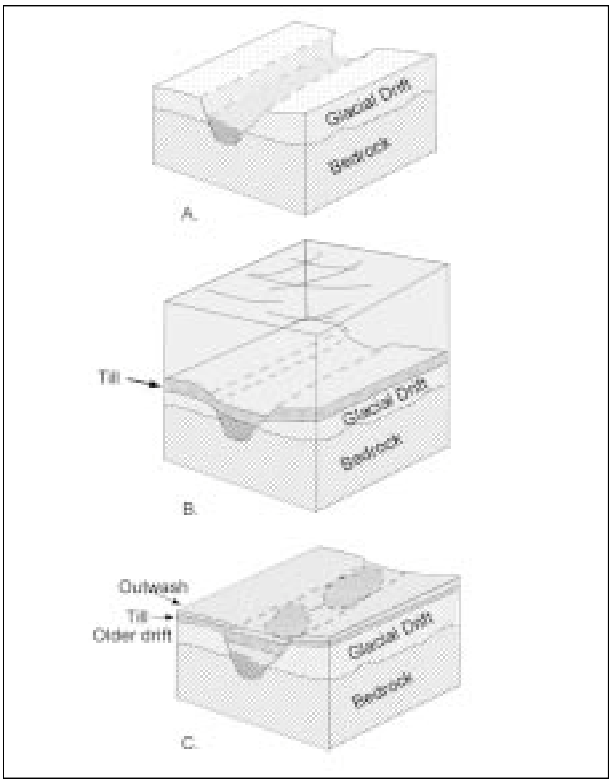


Fig. 7. Origin of Type III tunnel channels. A. Erosion of tunnel channel by Saginaw Lobe as in Fig. 5A and meltout of most or all of buried ice as in Fig. 5C. B. Readvance of Saginaw Lobe over site with deposition of diamicton (till). C. Outwash deposited during retreat of Saginaw Lobe or advance of Lake Michigan Lobe from west. Final valley is subtle, low relief topographic feature identified by linear chain of lakes on landscape.

Type IV Tunnel channels

The Type IV tunnel channel model is shown in Figs. 4B, 8 and 9A. This type of palimpsest tunnel channel forms when sediment from a different lobe than the lobe of origin partially buries the channel. The example in Figs. 4B and 9A was previously described by Kehew et al. (1999, 2005). As the Lake Michigan Lobe advanced into terrain that had been recently deglaciated by the Saginaw Lobe, large glaciofluvial fans were deposited from the ice margin shown in Figs. 4B and 9A. These fans slope toward the large Saginaw Lobe tunnel channel and continue with exactly the same slope on southeast side of the channel. The best way to explain this cross cutting relationship is that the tunnel channel was completely filled with ice and debris at the time of fan deposition. After the fans were deposited, the buried ice gradually melted and the modern valley formed.

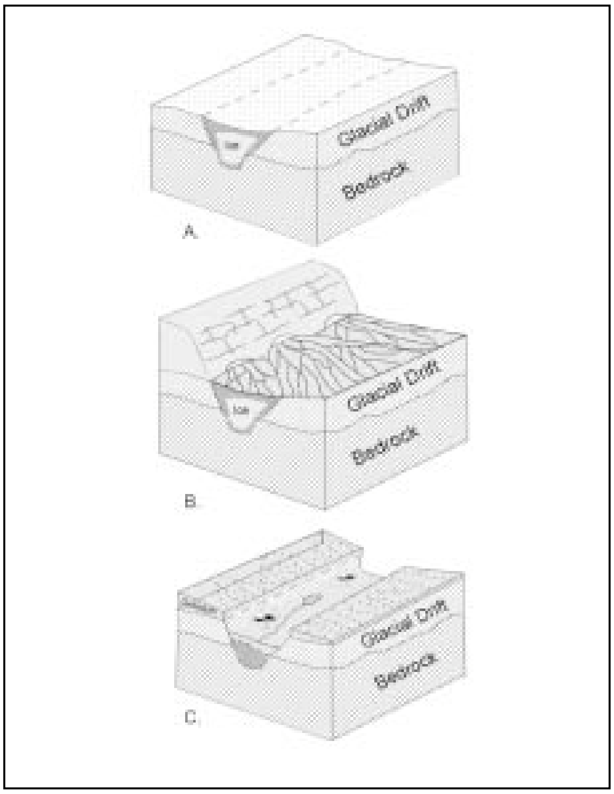
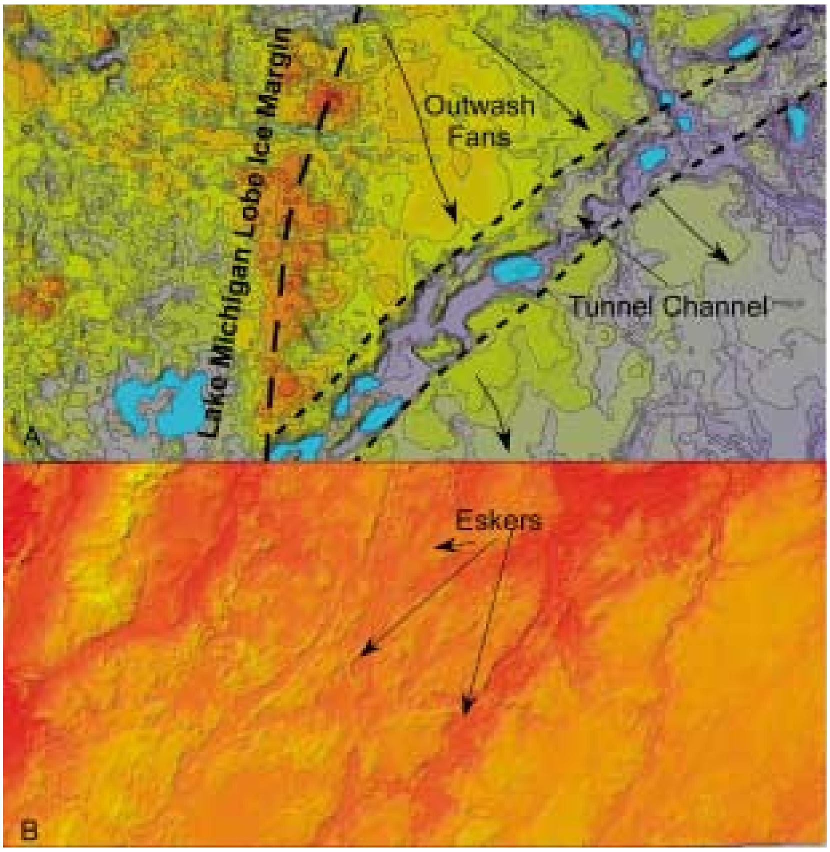


Fig. 8. Origin of Type IV tunnel channels. A. Tunnel channel is subglacially eroded and completely filled with ice and debris. B. Lake Michigan Lobe advances from west and deposits outwash fans over buried tunnel channel. C. Buried ice melts to form present-day valley. Slope of fan remnants west of valley slope is identical to slope of fan remnants east of valley.

Fig. 9. A. Digital elevation model of Type IV tunnel channel. Origin of channel conforms to model shown in Fig. 8. B. Digital elevation model of Type V tunnel channels containing eskers. Origin shown in Fig. 10.



Type V Tunnel channels

The last type of tunnel channel to be described is shown in Figs. 4A, 9B and 10. The presence of eskers in the valley bottoms of these channels indicates that they were not filled with ice and debris after channel erosion. An esker in one of these valleys was drilled using rotasonic methods in the summer of 2006 (unpublished data). The stratigraphy of the esker consists of a fining upward sequence from gravel in the tunnel channel to interbedded fine sand and silt near the top of the esker. The base of the tunnel channel was not reached in the borehole. This stratigraphic sequence indicates that flow in the channel declined from high velocities at the base to intermittent, low-energy conditions at the top. The tunnel may have contained impounded to very slowly moving water at times. These conditions suggest that the tunnel was occupied continuously or intermittently over a relatively long period of time, perhaps during stagnation and/or deglaciation of the lobe.

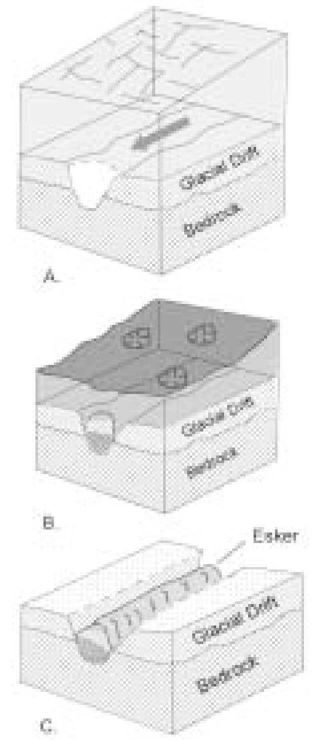


Fig. 10. Origin of Type V tunnel channels. A. Tunnel channel is subglacially eroded. B. Channel persists for long period of time during ice stagnation or retreat. Discharge and velocity through tunnel decreases through time. C. After retreat, esker with fining upward stratigraphic sequence remains in valley.

CONCLUSIONS

Tunnel channels of the Saginaw Lobe of the Laurentide Ice Sheet provide important clues to the subglacial hydrology of the lobe and the interaction of the lobe with its adjacent lobes. Five types of tunnel channels are recognized in this study:

- Type I–large, mostly unburied channels.
- Type II–partially buried palimpsest channels cutting an end moraine produced by readvance of same lobe.
- Type III–deeply buried palimpsest channels behind end moraines produced during readvance of same lobe.
- Type IV–palimpsest channels partially buried by deposits from a different lobe than the lobe under which they were formed.
- Type V–unburied to partially buried channels that contain eskers.

Two predominant sets of channels include southwesterly and westerly trending segments. The southwesterly channels appear to be the oldest, perhaps dating from the LGM or retreat of the lobe following the LGM. South of the Kalamazoo Valley, the tunnel channels are palimpsest features overridden or buried by sediment from more recent advances of the Saginaw or Lake Michigan Lobes. The presence of eskers in tunnel channels north of the Kalamazoo Valley suggests the persistence of these subglacial channels until late in the deglacial history of the lobe. Some of these channels were probably cut time-transgressively during the younger advance of the lobe to the Sturgis Moraine or more recent ice-marginal positions. The west-trending channels cross cut the southwesterly channels and are interpreted to represent the most recent episode of subglacial drainage. The change in orientation of these channels may be the result of retreat of the Lake Michigan Lobe, which was impinging upon the western margin of the Saginaw Lobe. Large, probably catastrophic subaerial meltwater flows can be traced from the west trending tunnel channels southward between the two lobes as the Lake Michigan lobe was retreating (Kozlowski et al. 2005). Further, more detailed study of these channels will be necessary to fully understand their origin and temporal relationships.

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Geological Survey of Finland, Special Paper 46, 79–86, 2007.

PROGLACIAL SEDIMENTARY ENVIRONMENT IN PÄRNU AREA, WESTERN ESTONIA AS DERIVED FROM THE VARVED CLAY STUDIES

by
Tiit Hang, Peeter Talviste, Raul Reinson and Marko Kohv[1]

Hang, T., Talviste, P., Reinson, R. & **Kohv, M. 2007.** Proglacial sedimentary environment in Pärnu area, western Estonia as derived from the varved clay studies. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 79–86, 5 figures, 1 table.

Distribution, grain-size and varve thickness data of the Late Weichselian varved clays have been analysed to describe the sedimentary environment and the ice sheet recession at the distal position of the Pandivere–Neva (13,500–13,300 yrs BP) ice-marginal formations, western Estonia. Dominating clay and silt fractions and very low sand content reflects an inflow of mainly fine-grained material. Vertical changes in grain-size are characterised by the decrease of clay content and increase of silt content towards the depth. Gradually decreasing upwards total varve thickness and characteristic changes in relation to seasonal layer thicknesses within each varve reflect normal sedimentation in the proglacial lake where, at each point, proximal conditions gradually changed to more distal conditions. There is no evidence from the varve clay studies pointing to the stagnation or readvance of the ice terminus prior or during the formation of the Pandivere–Neva ice marginal formations.

Key words (GeoRef Thesaurus, AGI): glaciolacustrine sedimentation, clay, varves, grain size, thickness, deglaciation, Pleistocene, Weichselian, Pärnu, Estonia.

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INTRODUCTION

The decay of Weichselian ice from Estonian territory took place between 14.7 and 12.7 ka BP (Kalm 2006). More often, compared to the preceding deglaciation stages, extensive proglacial bodies of water were present in front of the receding ice margin. This is reflected in the wide distribution of the glaciolacustrine sediments including varved clays. Glacial varved clays with their characteristic summer (silty) and winter (clayey) layers are interpreted to reflect seasonal variations in the sedimentation environment in the proglacial lake (De Geer 1940). Varve thickness changes, together with inside structure and fabric, can produce more details about the proglacial environment and the deglaciation. Interrupted chains of end moraines and glaciofluvial formations in Estonia mark marginal positions of the ice sheet in present topography. It is believed that they represent temporary stagnations (Karukäpp & Raukas 1997, Boulton et al. 2001) or even readvances (Raukas 1992, Raukas et al. 2004, Kalm 2006) of the ice terminus when the glacier regime was close to equilibrium. This, in turn, is most likely reflected also in the proglacial sedimentary environment in front of the ice margin. In the current article, the distribution of varved clays, grain-size and varve thickness changes are analyzed to describe the characteristics of the ice recession prior to the formation of the Pandivere–Neva belt of ice marginal formations in the Pärnu area, western Estonia.

GEOLOGICAL SETTING

The study area (Fig. 1) includes the coastal area south of the Pandivere–Neva (13,500–13,300 ka yrs BP., Hang 2001, Kalm 2006) belt of ice marginal formations in western Estonia. The surface of Devonian sandstones in the area lies at an altitude of -10 – -15 m above the present sea level (a.s.l.) (Tavast & Raukas 1982). Over the entire study area, the area is covered by bluish-grey loamy till of Late Weichselian age. The latter is followed by up to 30 m thick glaciolacustrine varved clay and silt. The thickness of Holocene marine sands and silts, covering these deposits, reaches at its maximum of 10 m, being mostly 2–3 m and is only locally absent.

Varved clays in western Estonia have been deposited in the Baltic Ice Lake (Fig. 1). These glacial clays are characterised by an average thickness of 10–20 m with very distinct laminations. Seasonal layers are easily distinguishable. Varves are usually thick and clayey (having the highest clay content 70–80 % among the Estonian glaciolacustrine clays) with silty microlayers in some sequences (Saarse & Pirrus 1988). Current ripple marks and erosional features are quite rare as well as synsedimentary disturbances.

The Pandivere–Neva ice marginal zone in Estonia starts from the West Estonian Archipelago and continues on the mainland towards the northeast (Fig. 1). Most often, the corresponding zone in northwestern Russia is considered to be the Neva zone (Raukas et al. 1971, Lundqvist & Saarnisto 1995, Raukas et al. 2004). The Pandivere–Neva zone is represented by a curved belt of push end-moraines, glaciofluvial deltas and marginal eskers (Karukäpp & Raukas 1997). In western Estonia the marginal eskers, which represent a specific type of glaciofluvial delta (Raukas 1992), are the most outlined landforms due to a more or less continuous length of 150 km. Due to Holocene wave erosion, the forms have been levelled while the height of features ranges only from a few meters up to a maximum of 20 m. In the Pärnu area, the Pandivere–Neva formations are buried under younger marine sediments.

Fig. 1. Location of the study area in western Estonia (green square). The distribution map of glacial varved clays in Estonia after Saarse & Pirrus (1998) with complements. Ice marginal positions are incorporated from Raukas et al. (2004).

METHODS

For the varve thickness studies and description of facial units of varved clays, 15 undisturbed sequences of varved clay were obtained with a Russian type peat corer in the surroundings of Pärnu and in the bottom of Pärnu Bay in the Baltic Sea (Fig. 2). Varve thickness study and construction of varve graphs followed the traditional method introduced by De Geer (1940). The varve graphs display the number of varves on the x-axes starting traditionally with the oldest varve on the right and the thickness of individual varve on the y-axes. The relative timescale is different for each varve graph.

Within the current study, a total of 84 samples, taken from three sediment cores, were analysed for the grain-size composition. The conventional pipette method was used for the silt and clay fractions (>0.063mm; 0.063–0.016mm; 0.016–0.008mm; 0.008–0.002mm; <0.002mm), while wet–sieving was done for the coarse (>0.063mm) material. In addition, earlier results of grain-size analysis of 105 samples from 11 geotechnical corings were used to describe the grain-size distribution of glacial varved clay in the Pärnu area.

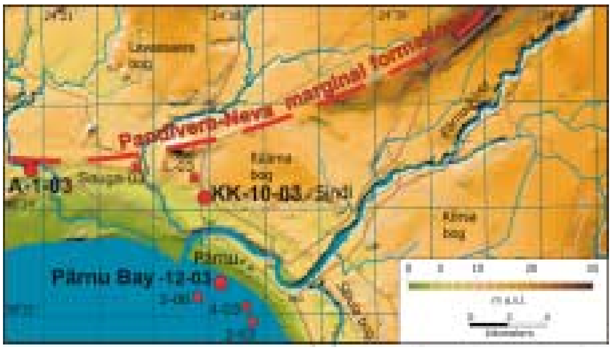

Fig. 2. Location of the studied sequences (circles) and clay sequences discussed in the text (squares) at the distal position of the Pandivere–Neva ice marginal formations in western Estonia. At the Audru site (A–1–03), 8 parallel sequences were studied for varve thickness changes.

Morphology of the upper surface as well as thickness and distribution of clay is described through data originating from 848 geotechnical corings from the Pärnu area.

RESULTS

Distribution of varved clays in Pärnu area

Over the entire study area, the hummocky upper surface of the Late Weichselian bluish-grey loamy till is covered by glaciolacustrine varved clay or silt (Fig. 1). Glaciofluvial deposits on top of the till have been only locally reported. The maximum thickness of varved clay in the Pärnu area is 30 m, but on average between 5–9 m. Locally rapid changes in clay thickness are due to the underlying topography. The upper surface of the clay dips towards the south at an altitude of –8 m a.s.l. approximately 5 km offshore and –2.5/–3 m a.s.l. at the current coastline of Pärnu Bay. At the distance of ca. 2 km north of coastline, the altitude of the clay surface rapidly changes from zero to 4 m a.s.l..

Grain-size

Clay and silt fraction dominate in the grain-size composition of proglacial clays in the Pärnu area (Fig. 3). With few exceptions, a common clay texture exceeding 60% is typical for the winter layers through the studied sections (Fig. 3A). Data points expressing the texture of the summer layers is grouped towards the textural class of silt but is more scattered compared to the data of the winter layers (Fig. 3A). Currently analysed massive diamicton has the coarsest grain-size plotted in Fig. 3A. Earlier grain-size analysis of mean samples (different seasonal layers included in one sample) support the current results (Fig. 3B) as the samples are closely grouped within the textural classes of silt and clay. It is obvious that the analysis of annual or multi-annual samples clearly underestimates the true content of clay and silt fraction in the deposit (Fig. 3). However, the analysed section PB-12-03 (Fig. 3A) probably influences the results as only this section contains clay unit E. Results of multi-annual sample analysis (Fig. 3B) clearly demonstrate vertical changes in grain-size composition (Fig. 3B) as the content of silt fraction increases and the clay

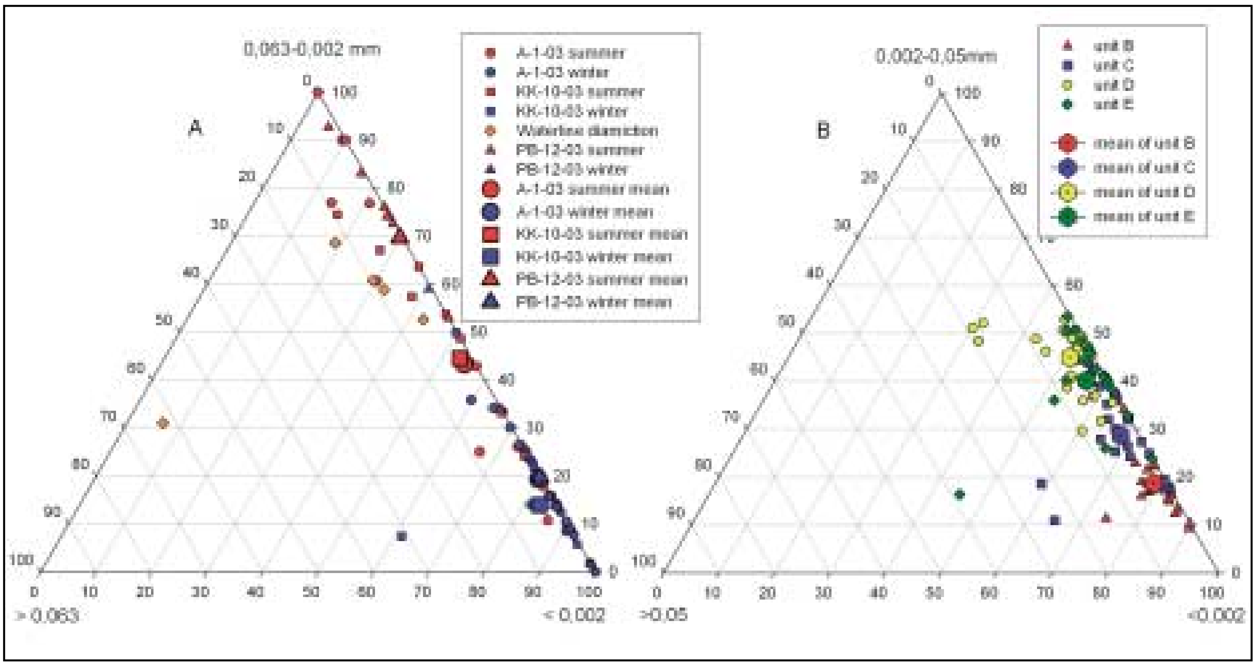

Fig. 3. Triangular plots of the grain–size composition of studied sequences (A) and clay units discussed in the text (B). Analysed summer and winter layers (A) belong to the same varves. Note, the different scales on the plots. Location of the studied sequences see in Fig. 2.

fraction decreases towards the depth (from unit B to unit E). Unit D, representing the aforementioned clayey diamicton, contains the coarsest material. In Fig. 4A, details about vertical changes in grain-size composition of sequence KK–10–03 (Fig. 2) are displayed. An increase of clay fraction and a decrease of fine-silt (0.008–0.002 mm) upwards are the main tendencies in summer layer composition. Notable changes in summer layers are taking place in the lower and upper portion of varved clay sequences both characterized by higher percentage of silt fraction (Fig. 4A). Grain-size composition of the winter layers is highly dominated by the clay fraction that in the lower portion of clay constitutes about 50–60% and in the upper part remains over 80%. Contents of the fine-silt fraction ranges between 5–10%. The content is higher (up to 35%) in the lowermost and uppermost portion of the varved clay (Fig. 4A).

Eight samples were analysed within an individual varve for their grain-size composition (Fig. 4B). Fine- and medium-silt fractions are predominant in the summer layer. Contents of the clay fraction increase rapidly in the upper part of the summer layer and begin to dominate at the transition from the summer to the winter layer (Fig. 4B). Somewhat surprising is the increased value of the coarse-silt (0.063–0.016 mm) fraction content in the upper portion of the summer layer.

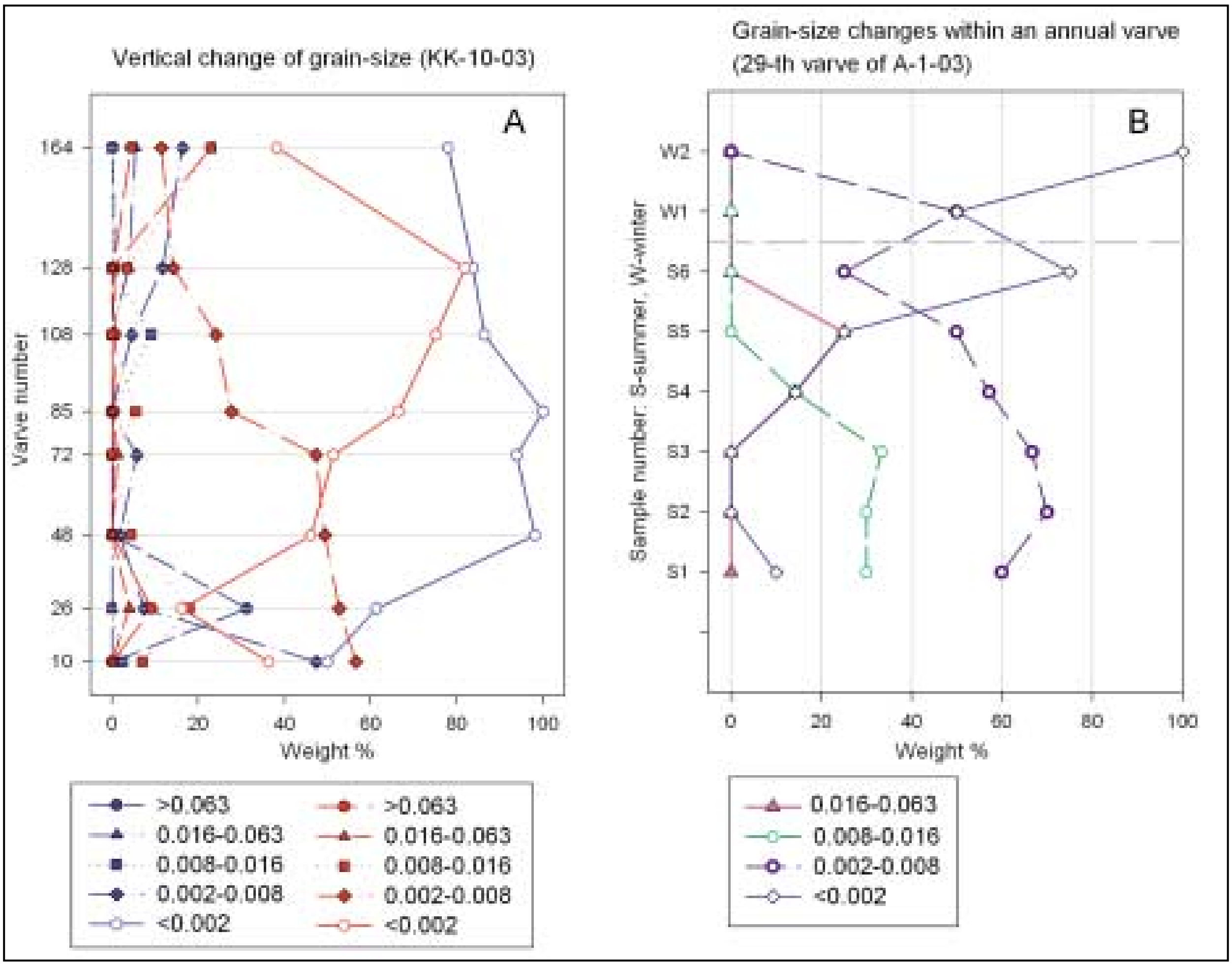

Fig. 4. Vertical changes of the grain-size composition in the varved clay sequence (A) and within an individual varve (B). On plot A, the red colour indicates the summer layers and the blue colour indicates the winter layers. Horizontal dashed-line on plot B marks the visual boundary between the summer and the winter layer. Location of the sequences seen in Fig. 2 and vertical location of the analysed varves within the sequences seen in Fig. 5.

Varve thickness changes

Varve thickness changes in three clay sequences from different key areas are displayed in Figure 5. Varve graphs are presented on a similar scale to the background of the relative timescale with the zero different for each sequence. According to total varve thickness changes and the relation between seasonal layer thicknesses within each varve, the varve series are divided into three groups (E, C, B). Summer layers dominate (52–80% of individual varve thickness) within the thick varves of group E (Fig. 5). Greenish grey silty summer layers display alternating silt and sandy silt layers or in places ripple bedding. Winter layers consist of reddish brown massive clay. Seasonal layers are easy to distinguish.

Winter layer thickness dominates (51–85% of individual varve thickness) within the varves of group C (Fig. 5). Although the individual layer thicknesses changing over a great range, the overall tendency is towards an increase in relative winter layer thicknesses. Total varve thickness decreases upwards. Greenish grey massive silt constitutes the summer layers and dark brown massive clay the winter layer. In some varve intervals within group C, distinguishing between the seasonal layers is problematic.

The third group of varves (B) is characterised by decreased thickness compared to the previous varve groups. The thickness of individual varves changes greatly as do the seasonal layer thicknesses within an individual varve (Fig. 5). Distinguishing between the seasonal layers is problematic. In the upper portion of this group, visually distinguishing between the varves is also problematic and this is why the number of varves in this group is usually bigger than displayed in any of the varve graphs.

Only the varves of group E with clear dominance of the summer layers are present in the sequence from Pärnu Bay (Pärnu Bay-12-03) (Fig. 5C). Erosional discontinuity at the upper contact with overlying marine sand points to the post-sedimentary erosion at this location.

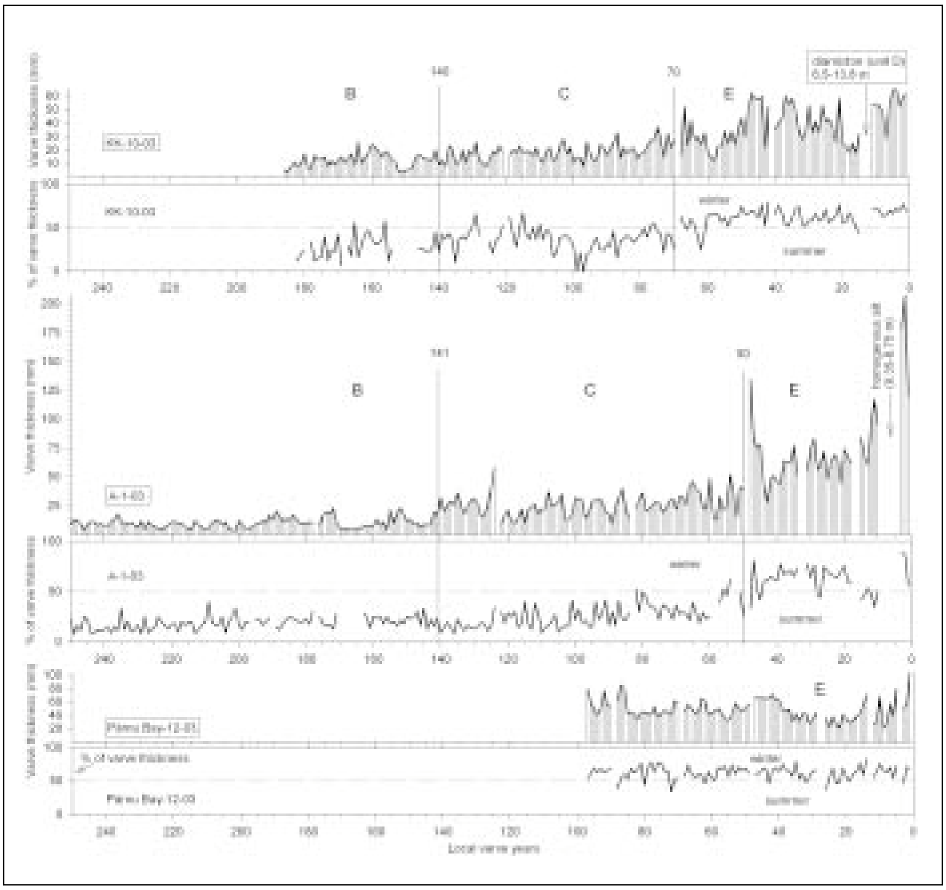

Fig. 5. Examples of the varve thickness graphs of glacial varved clays from the sediment sequences of the Pärnu basin. Lower diagrams on each plot show the relation of individual layer thickness within a varve. E, C, B varve groups of different clay units discussed in the text. Vertical lines, indicating a certain varve, point to the transition between the varve groups. Note that the graphs on the figure are not correlated between each other. Location of the sequences is displayed in Fig. 2.

DISCUSSION AND CONCLUSIONS

According to different geotechnical parameters (mineralogical and chemical composition, plasticity, water content) (Kattel 1989) as well as varve thickness, inside structure and fabric, the glacial clay complex in the Pärnu area could be divided into five facial units (A-E) (Table 1) with transitional rather than fixed boundaries. Dehydrated clay in unit A has been recorded at a higher (more than 5 m) altitude and represents up to 1 m thick upper clay interval that emerged due to water level lowering after the clay accumulation. Periodic freezing-melting cycles and accompanying chemical processes have destroyed the varve structure, and where present, the unclear thin lamination has been reported. Unit B represents a series of thin winter layer dominating clayey varves with changing thickness between individual couplets. These uppermost varves are interpreted to have been formed in the distal part of the proglacial lake. Moving downwards, unit C varves are also winter layer dominating clayey couplets. Varve thickness is greater compared to the previous interval. Thickness of clearly distinguishable individual couplets increases downwards. The clay of unit C is interpreted to have been formed distally far from the areas of sediment influx, in the deeper central part of the proglacial lake. Thickest varves of the clay sequences in the Pärnu area contain the coarsest material. Multiple graded summer layers dominate in these lowermost varves, which belong to the unit E and are formed in varying sedimentary conditions close to the ice margin. Normal varve series from the proximal varves (unit E) to distal varves (units A, B) is interrupted by an interval of waterline glacial diamicton (unit D). Unit D represents an up to 8 m thick massive silty-clay with dispersed grains of sand. This unit has a scattered distribution with maximum thicknesses in the northern and northeastern part of the study area and is missing in the nearshore area of Pärnu Bay. Similar deposits have been described earlier in the West Estonian Archipelago and are interpreted as a waterline diamicton (Kadastik & Kalm 1998). The scattered distribution of unit D in Pärnu area is hard to explain. The selected erosion is most likely not an option while in clay sequences from Pärnu Bay, an interval of ca. 10 varves with dispersed grains of sand in the winter layers were reported. According to preliminary varve correlation, the above interval stratigraphically corresponds to unit D described in the sequences from the city of Pärnu and demonstrating, thus, that the accumulation of unit D deposits is the result of one sedimentary event in the proglacial environment. Unfortunately, details about the genesis of this unit currently are unclear.

Dominating clay and silt fraction (Fig. 3) and low (2–5%) sand content reflects an inflow of mainly fine grained material. Continuously increasing clay and decreasing silt content upwards demonstrate that the proximal sedimentary conditions at each point in the proglacial lake gradually changed to become more distal. This conclusion is also supported by the thickness changes of the varves. All investigated sequences display a normal varve series with decreasing varve thickness upwards. From studies into grain-size, internal structure and thickness changes of the varves, there is no evidence to conclude any stagnation and/or readvances of the ice margin during the varve formation. This is surprising because the study area is located at the distal position from the Pandivere–Neva ice marginal formations,

Table 1. Facial units distinguished in the proglacial clay sequences in the Pärnu area, western Estonia.

Facial units	Description
A	Dehydrated greenish-grey clay, varve structure destroyed due to emergence after clay accumulation and due to soil forming processes
B	Varved clay containing a series of thin winter layer dominating clayey, hardly distinguishable seasonal couplets
C	Varved clay, varves are winter layer dominating with clearly distinguishable clayey seasonal couplets; total varve thickness is less than in unit E and its decreasing upwards. Where present, the lower boundary with unit D is sharp
D	Waterline glacial diamicton; grey massive silty-clay with dispersed grains of sand; discontinuous lateral distribution; maximum known thickness 8 m
E	Varved clay; summer layer dominating varves with clearly distinguishable seasonal layers; thickest varves of clay sequence with upwards decreasing varve thickness; unstable sedimentary environment close to the ice margin is reflected in multiple graded summer layers with rare ripples; where present the upper boundary with unit D sharp otherwise the transition from unit E to C displays continuous lamination

which is believed to have formed during the temporary stagnation of the ice terminus. Moreover, the normal series of varves, with only some disturbed intervals, were also reported at a distance less than one kilometre from the marginal formations. This highlights the question of the formation of the Pandivere–Neva belt of ice marginal formations. For further conclusions, additional lithostratigraphical and sedimentological as well as varve correlation are needed across the Pandivere–Neva formations.

In summary, the following conclusions can be drawn:

– Clay complex in the Pärnu area is divided into five facial units (A-E) with transitional boundaries.
– Units E and C-A represent a normal series from proximal to distal varves. Upper dehydrated clay unit A lost its laminated character due to the emergence after the clay accumulation. Unit D represents a waterline glacial diamicton with a limited distribution and variegating thickness. Details about the genesis of this unit are unclear.
– Dominating clay and silt fraction and very low content of sand reflects an inflow of mainly fine-grained material.
– Grain-size changes within an individual varve demonstrate the dominance of the fine to medium grained silt fraction in the summer layer, very rapid increase of clay content below the upper boundary of the summer layer and clear dominance of the clay fraction in the winter layer. Rapid increase of coarse-silt content at the end of summer season could be connected to increased storm activity before the cold season.
– According to total varve thickness, the relation between thicknesses of seasonal layers, internal structure and fabric of the varves, a continuous sedimentation during gradual ice recession was determined with a clearly defined series from proximal to distal varves.
– From the varve clay studies, there is no evidence pointing to the stagnation or readvance of the ice terminus prior or during the formation of the Pandivere–Neva ice marginal formations.

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BURIED ORGANIC SEDIMENTS IN ESTONIA RELATED TO THE ANCYLUS LAKE AND LITORINA SEA

by
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Saarse, L., Vassiljev, J., Miidel, A. & Niinemets, E. 2007. Buried organic sediments in Estonia related to the Ancylus Lake and Litorina Sea. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 87–92, 5 figures.

A database of buried organic deposits related to the development of the Ancylus Lake and Litorina Sea in Estonia was created. Of the 78 sites included in that database, 45 are of pre-Ancylus and Ancylus and 33 of post-Ancylus and Litorina age. Radiocarbon dates of buried organic beds have a large age scatter. In several cases, recent radiocarbon dates are older than those obtained earlier, referring to methodological problems or failure of sample collection and preservation. Buried organic deposits of Ancylus Lake age are mostly characterised by the dominance of Pinus pollen. This was the main reason why the age of the Ancylus transgression in Estonia was considered to be about a thousand years younger than in the neighbouring countries. The pollen spectra of the Litorina buried organic strata differ considerably between sites. A simulation of buried organic beds and shoreline isobases shows discrepancies, especially in the surroundings of Pärnu Bay, obviously due to landslides, re-deposition, erosion or erroneous initial data. Comparisons of shore displacement curves and the ancient settlement positions do not support the concept of multiple Litorina Sea transgressions.

Key words (GeoRef Thesaurus, AGI): buried soils, organic sediments, Ancylus Lake, Litorina Sea, pollen analysis, absolute age, ^{14}C, Holocene, Estonia.

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INTRODUCTION

The study of beach formations and littoral deposits in Estonia, extending back to ca 150 years, was initiated by Schmidt (1869) and followed by the classic works of Hausen (1913) and Ramsay (1929). These studies were advanced by the elevated position of shorelines due to the location of Estonia in the glacioisostatically uplifting area. The organic strata buried under the coastal formations of the Ancylus Lake and Litorina Sea was discovered for the first time on the banks of the Pärnu and Narva rivers (Hausen 1913, Thomson 1933, 1937). By the 1960s, ten sites with organic deposits buried during the Baltic Sea transgressions were known (Kessel 1960). The increasing use of gravel and sand deposits accelerated the exploration of resources, during which a number of new gravel pits were opened and many new buried peat sites were discovered. By 1968, H. Kessel had already described 38 such sites. These are mostly small compressed peat patches, less than 50 cm, occasionally up to 100 cm in thickness, embedded into sand and gravel (Veski et al. 2005). Especially in the Pärnu area, buried peat is up to 2.7 m thick (Talviste et al. 2006).

The present paper gives an overview of the distribution and age of the buried organic deposits, which were submerged during the Ancylus Lake and Litorina Sea transgressions, considering also the shore displacement, isobases and the position of buried deposits.

MATERIAL AND METHODS

The shoreline database compiled recently for the Baltic Ice Lake, Ancylus Lake and Litorina Sea contains more than 600 sites, 424 of which are from Estonia and the rest from adjoining areas (Saarse et al. 2003). The Holocene buried organic sediment database of Estonia includes 78 sites: 45 sites of pre-Ancylus and Ancylus age, and 33 sites of post-Ancylus and Litorina age. The database includes the site's number, name, elevation and coordinates, calibrated and uncalibrated radiocarbon dates and laboratory number, description of dated material, and references. The complete list of the shoreline database of Estonia and neighbouring areas is not yet available, but the Estonian database of the buried organic beds was recently published (Saarse et al. 2006).

The above-mentioned databases were used to reconstruct water-level surfaces of the Ancylus Lake and Litorina Sea using a point kriging interpolation with linear trend approaches (Saarse et al. 2003). Residuals (actual site elevation difference from the interpolated surfaces) for every site were calculated and sites with residuals of more than 0.7 m were discarded. The surfaces of the buried organic sediments were reconstructed in the same way for two time intervals: 10,800–9,500 cal BP (9,500–8,500 uncal BP) and 9,400–7,800 cal BP (8,400–7,000 uncal BP) and compared to the water-level surfaces of the Ancylus Lake and Litorina Sea, respectively.

RESULTS AND DISCUSSION

The buried organic beds are positioned along the north and west coasts of the Estonian mainland and on the islands of Hiiumaa and Saaremaa (Figs. 1a, b). The coastal formations of the Ancylus Lake are located at heights between 45 and 3 m a.s.l. (above the present day sea level), while the highest Litorina Sea limit lies between 25 and 5 m (Fig. 2). This means that near the Estonian/Latvian border, the Ancylus deposits lie at a lower altitude than the Litorina ones. The surface of the buried organic bed commonly remains 1–2 m lower than the highest shoreline limit. In northern and western Estonia, the

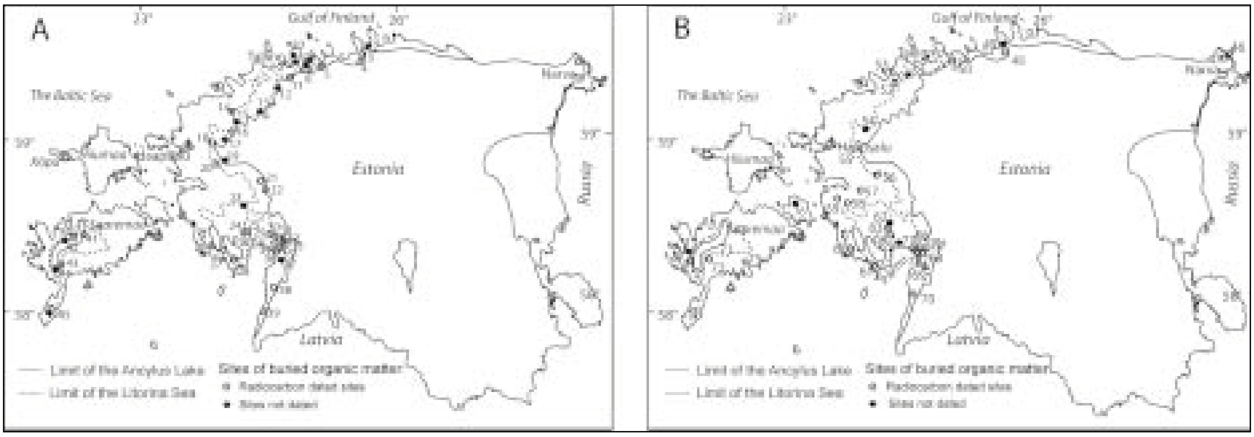

Fig. 1. Distribution of Holocene buried organic beds in Estonia. Organic deposits buried (A) during pre-Ancylus time and Ancylus transgressions, and (B) during post-Ancylus time and Litorina transgressions. Modified after Saarse et al. (2006).

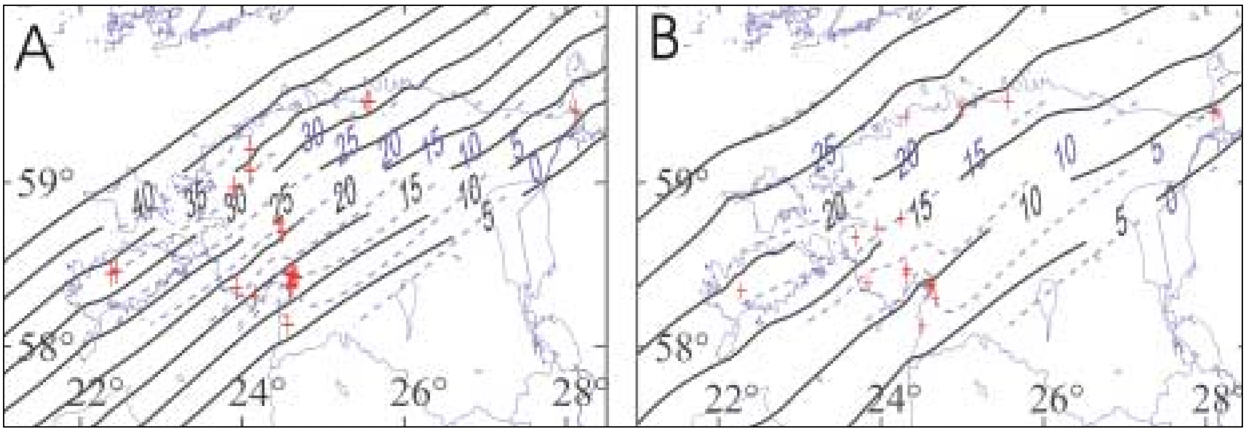

Fig. 2. (A) Isobases of the Ancylus Lake shore displacement (solid line) and buried organic deposits lower limit (dashed line) and (B) isobases of the Litorina Sea shore displacement (solid line) and buried organic deposits lower limit (dashed line). Isobases of organic matter show disturbances in the surroundings of Pärnu, discussed in the text. Red crosses (+) mark the sites with buried organic deposits used in modelling. Shore displacement isobases are modified after Saarse et al. (2003).

thickness of buried deposits is smaller than in the Narva and Pärnu areas, where the covering strata are up to 4–5 m thick.

Comparisons of the Ancylus Lake isobases with the isobases of buried organic matter (mainly peat) displays nonconformity in the Pärnu area (sites 35, 36, Figs. 1a, 2a). Both conventional and AMS radiocarbon dates verifies that peat was formed during the Ancylus transgression, but lies about 4–5 m lower than at the nearby sites. One can suspect that buried peat is not in situ position, as marine deposits rest on varved clays, which are highly susceptible to landslides. Similar nonconformity appears when comparing the isobases of the Litorina Sea shorelines with the isobases of buried organic matter. At some sites (Figs. 1b, 2b; sites 77, 78) in the Pärnu area, a great difference (up to 12 m) has been recorded in the elevation of the highest shoreline and the buried strata surfaces (Talviste et al. 2006), and the buried peat lies below the modern sea level. The reason for such phenomena is not yet clear and more detailed studies are required. However, the low position of the buried peat could be explained by low water levels prior the Litorina transgression (Talviste et al. 2006).

At present, 71 radiocarbon dates are available from the pre-Ancylus and Ancylus buried organic deposits of 30 different sites (Fig. 3). Uncalibrated ^{14}C radiocarbon dates vary between 9,980±120 (Tln-2349) and 8,440±70 (TA-263). Buried peat, which was formed during the Ancylus Lake regression (pre-Litorina beds) and Litorina transgression,

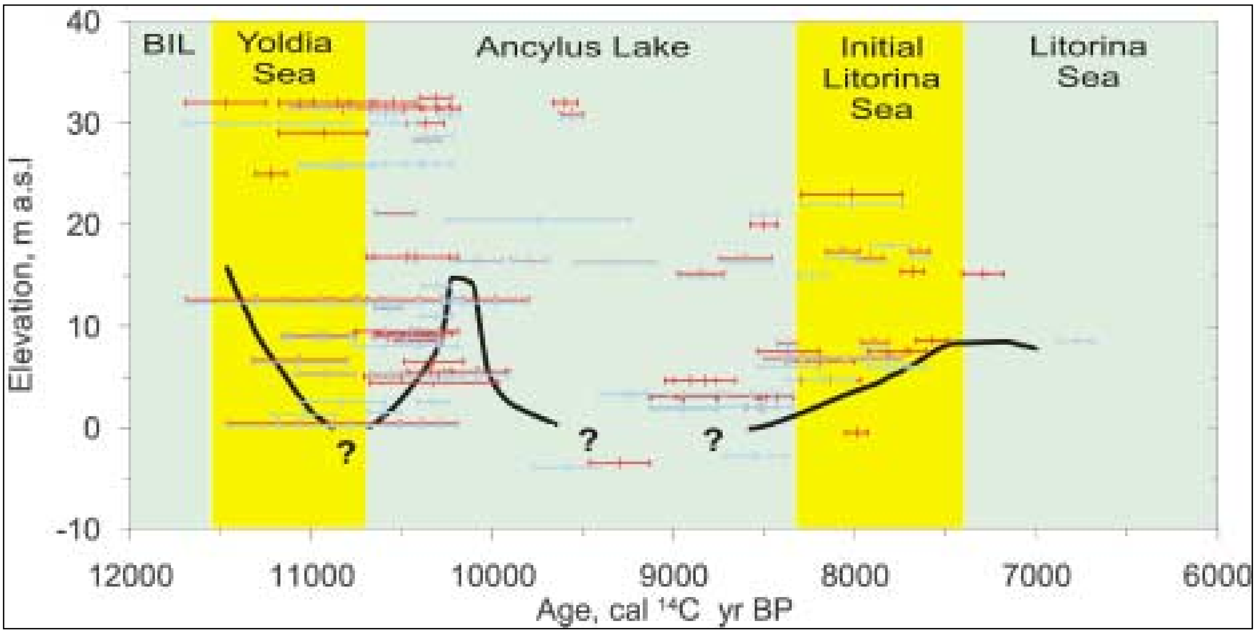

Fig. 3. Distribution of radiocarbon dates with error bars. The blue line marks the age of the lower limit of the buried organic beds (cal BP), red line – the age of the upper limit. The bold black line shows water level changes in the surroundings of Pärnu (after Veski et al. 2005). The age of the different phases of the Baltic Sea basin according to Andrén et al. (2000). (BIL = Baltic Ice Lake).

has been dated in 22 localities and a total of 66 ^{14}C dates are available (Fig. 3; Saarse et al. 2006, Talviste et al. 2006). Uncalibrated dates range from 8,570±150 (Ta-2843) to 5,520±100 (Tln-178). Obviously, this is due to different factors as radiocarbon dates depending on the material analysed (wood, seeds, bulk organic matter, insoluble or soluble fraction), its preservation and availability to weathering, contamination with older carbon or younger rootlets (Olsson 1986, Wohlfarth et al. 1998). Wood, peat and seeds from locations 27, 28 and 35 (Fig. 1a), all connected with the Ancylus Lake development, were subjected to conventional and AMS radiocarbon dating. It was determined that the dates from the peat were older than those from wood and seeds at the same depth (Veski 1998). Several sites (5, 11, 21, 31, 70; Figs. 1a, b) were re-examined. The new ^{14}C dates obtained are commonly older than the previous ones and more consistent with pollen stratigraphy (Veski 1998). One reason for incorrect dates could be the preservation of samples (not kept in a refrigerator) and the time span between the collection and dating, which in some cases was formerly more than one year.

According to the re-examined radiocarbon dates, the Ancylus Lake transgression started about 10,800 cal BP and culminated ca 300–500 years later. Around 10,100 cal BP, a rather rapid regression of up to 30 m (Raukas et al. 1996) followed in the areas of rapid uplift. Buried organic deposits of Ancylus age are mostly characterised by the maximum of Pinus pollen (Fig. 4). This was the main reason why the Ancylus transgression in Estonia was estimated to be about 1,000 years younger (Kessel & Raukas 1979) than in neighbouring countries. According to radiocarbon dates, the Litorina transgression started about 8,300–7,800 cal BP and culminated at different times in different regions, first in the areas of rapid uplift (Kessel & Raukas 1984, Miettinen 2002). Pollen spectra of the Litorina Sea buried organic strata differ considerably between sites, still showing an affiliation to the Atlantic chronozone.

Comparisons of shore displacement curves with the positions of ancient settlements (Fig. 5) do not support the idea of multiple Litorina transgressions. However, this does not rule out smaller sea-level changes caused, for example, by stormy waves. Detailed bio- and chronostratigraphic analyses are needed to study the character of Litorina transgression more precisely.

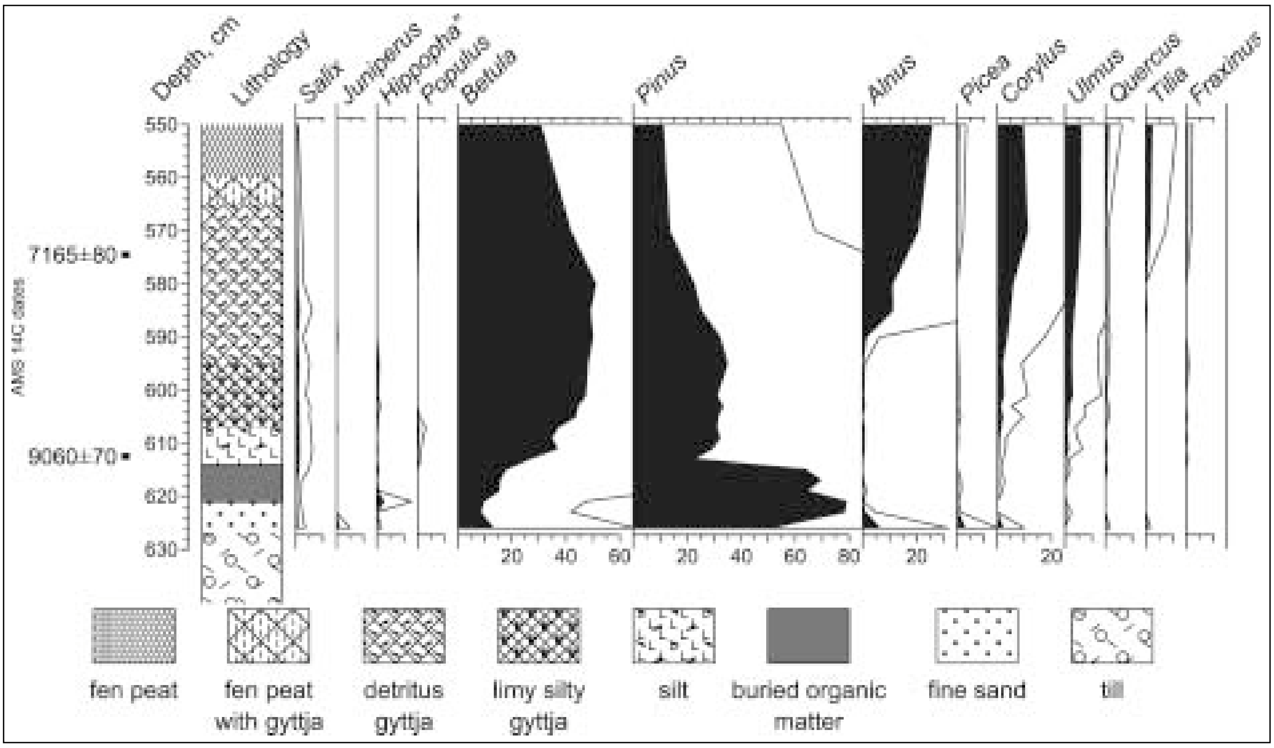

Fig. 4. Pollen diagram from Kõpu (No 28, Fig. 1a). Buried organic beds are enriched with Pinus pollen, which was the main reason to consider their formation during the Boreal period. Analyses by S. Veski.

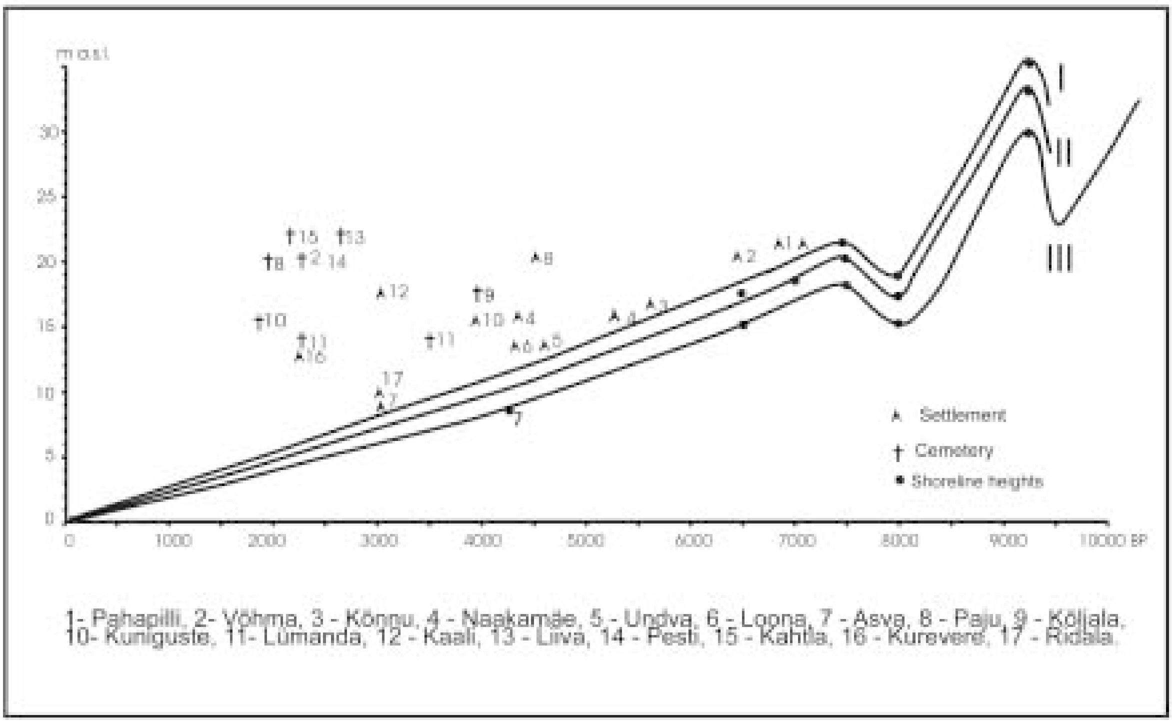

Fig. 5. Comparison of shore displacement curves from Saaremaa with the settlement sites and cemetery locations. Ages uncal BP. (I) shore displacement curve for site 35, (II) site 33 and (III) 30 m Ancylus isobase in Saaremaa Island.

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Geological Survey of Finland, Special Paper 46, 93–98, 2007.

MACROFOSSIL EVIDENCE DISPUTE UBIQUITOUS BIRCH-PINE-SPRUCE SUCCESSION IN WESTERN FINNISH LAPLAND

by
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Sutinen, R., Aro, I., Herva, H., Muurinen, T., Piekkari, M. & Timonen, M. 2007. Macrofossil evidence dispute ubiquitous birch-pine-spruce succession in western Finnish Lapland. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 93–98, 5 figures.

Reconstruction of the post-glacial migration of Norway spruce (*Picea abies (L.) Karst.*) into northern Fennoscandia is based on pollen records and radiocarbon (^{14}C) ages from organic deposits. However, no macrofossil evidence indicates the distribution of spruce beyond its present timberline-treeline ecotone. We applied total sampling to study Holocene succession of tree species at two sites: Pousujärvi peat bog (68°51'N, 21°10'E), beyond the present tree limits of spruce and Scots pine (*Pinus sylvestris L.*) and at Lake Kompsiotievanlammit (68°30'N, 22°30'E) at the present pine treeline in western Finnish Lapland. Subfossil pine logs at the Pousujärvi site yielded ^{14}C-ages from 5,000±40 to 5,110±60 yr BP (5,730 to 5,900 cal. BP). The pioneer species were found to be birch (*Betula pubescens ssp.*) yielding ^{14}C-age of 7,980±60 yr BP (8,810 cal. yr BP). No spruce was present. All subfossil trunks sampled from Lake Kompsiotievanlammit were pines, and yielded ages less than 3500 yr BP. No spruce was present. Our results imply that Holocene tree species succession is incomplete in that spruce is absent in the chrono-stratigraphic sequences in regions where an edaphic dispersal barrier for spruce is comprised of tills derived from felsic rocks, particularly granulites.

Key words (GeoRef Thesaurus, AGI): paleobotany, *Betula, Pinus, Picea*, wood, mires, lakes, absolute age, ^{14}C, Holocene, Lapland Province, Finland.

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INTRODUCTION

The static model of the (aspen-) birch-pine-spruce successional sequence advocates that forests repeat the same regeneration and succession stages after forest disturbances (Pastor et al. 1999) and within the Holocene migration of tree species. Conifer tree limits and timberlines, however, display evidence of diverse succession and migration patterns during the Holocene. East of Fennoscandia, spruce (Picea abies ssp. obovata) forms the present coniferous tree limit reaching 66°20'-69°25'N, whereas the tree limit of Scots pine (*Pinus sylvestris L.*) remains at latitudes 64°30'-66°N, respectively (Veijola 1998, MacDonald et al. 2000, Payette et al. 2002). In Fennoscandia, the northern limits of Norway spruce (*Picea abies (L.) Karst.*) and Scots pine both reach a latitude of 70°N, but the timberline of spruce is located at significantly lower latitudes than Scots pine (Kallio et al. 1971, Veijola 1998, MacDonald et al. 2000; Fig. 1). Even though spruce is a climax species in the successional sequence, it is absent over a large region in the northern boreal forests of Lapland (Fig. 1), where Scots pine is a dominant species and birch (*Betula pendula Roth.* and *B. pubescens Ehrh.*) is a secondary species.

Holocene migration of tree species, tree-line fluctuations and forest disturbances are associated with climate-driven factors (Moe 1970, Eronen & Hyvärinen 1987, Kremenetski et al. 1999, MacDonald et al. 2000, Kullman 2002, Payette et al. 2002). Evidenced by ^{14}C-dated subfossil trunks, the expansion of spruce to sites north of the modern treeline in Eurasian Russia occurred 9,000-8,000 yr BP, and the retreat of spruce to its current treeline position occurred 4,000-3,000 yr BP (Kremenetski et al. 1998, MacDonald et al. 2000). Also, spruce macrofossils well above the present alpine tree limit in the Caledonian mountain range in central Sweden indicate that spruce was part of the forest migration during the early Holocene time period (8,000-11,000 yr BP; Kullman 2002). Hence, climate factors did not constrain the early migration of spruce in Fennoscandia. However, the conceptual pollen-derived migration model advocates that spruce reached its northernmost range limit as late as 3,500 yr BP in northern Fennoscandia (Moe 1970, Kremenetski et al. 1999). We argue that lithological provinces constitute edaphic dispersal barriers, such that spruce is absent on terrains with low soil water availability and nutrients and typified by acidic tills derived from felsic rocks, such as the granulite belt in northeast Fennoscandia (Sutinen et al. 2002, 2005). We attempted to see if spruce had been present, similar to the Russian (MacDonald et al. 2000) or Swedish (Kullman 2002) cases, in the early Holocene succession beyond its present tree limit/ timberline.

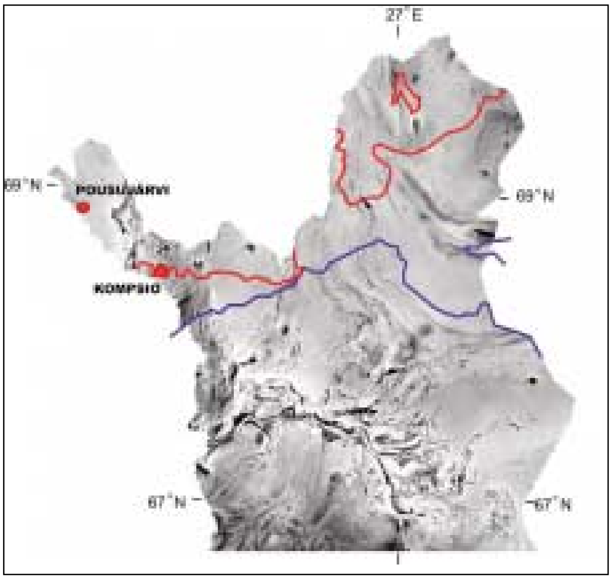

Fig. 1. Study sites in northern Finland. Timberlines of Norway spruce (blue line) and Scots pine (red line) adopted from Sutinen et al. (2005).

MATERIALS AND METHODS

Study sites

The Pousujärvi (68°51'N, 21°10'E) site was selected because one of the present co-writers (M. Piekkari) discovered (within the geochemistry mapping of tills by the Geological Survey of Finland) that subfossil logs were present in peat layers. The Pousujärvi site is located beyond the present spruce and pine (*Pinus silvestris L*) tree limits (Fig. 1). The Lake Kompsiotievanlammit (68°30'N, 22°30'E; Fig. 1) site was selected on the basis of relative location to the present Scots pine treeline. Previous studies by Eronen et al. (2002) have indicated that subfossil pine logs are common in the lakes of Lapland, hence we located potential survey sites with airborne surveillance. A helicopter search revealed that logs were present on the bottom of Kompsiotievanlammit.

Sampling

Both selected sites were logistically accessible, hence we applied a total megafossil sampling to verify a tree species succession pattern in the outermost corner in northwest of Finnish Lapland. We used a 20 ton excavator at the Pousujärvi site to sample all subfossil trees 2.5 m in thickness from the peat bog. At the Kompsiotievanlammit site, we used haul-skidder with a 12-m-long loading crane and log grips (Fig. 2) to sample all subfossil trunks at the bottom of the lake. At both sites, disk samples were collected with a chain saw (Fig. 3). The peat profile was investigated and sampled for ^{14}C-datings at the Geological Survey of Finland. Disks from the subfossil logs were analysed at the dendrological laboratory at the Finnish Forest Research Institute, Kolari station.

Fig. 2. Sampling with loading crane and log grips.

Fig. 3. Disk sampling for ^{14}C-datings and tree-ring analyses.

RESULTS AND DISCUSSION

Birch megafossils

Near the bottom of the bog at the Pousujärvi site a birch (*Betula pubencens ssp.*) layer was detected. The most distinctive subfossil birch trunk, 20 cm in diameter and from 220cm depth (Fig. 4) yielded a ^{14}C-age of 7,980±60 (8,810 cal. BP), making this the oldest tree megafossil found in the Kilpisjärvi area, Finland. The supra-long Scots pine tree-ring record by Eronen et al. (2002) extends to 7,500 yr BP, but the oldest ^{14}C-dated pine log in Lapland had yielded a date of 6,930 yr BP (7,680 cal. BP) (Eronen 1979). Wood material from the birch trunks was decomposed, but perfect in shape such that white bark and tree-rings were clearly visible (Fig. 4). Birch megafossils are not commonly found in Lapland, but one has been found beneath paleo-landslide materials in Kittilä (67°47'N, 25°28'E) yielding a ^{14}C-age of 8,720 yr BP (9,730 cal. BP) (Sutinen 2005). Pollen statigraphy at the Toskaljavri site (69°11'N, 21°27'E at 704 m a.s.l.) (Seppä et al. 2002) also indicates that birch was the pioneer tree species after the deglaciation of the Kilpisjärvi area, but the time of early migration of 9,600 cal. yr BP is not concurrent with the megafossil data obtained here. Our observations are also supported by the findings of birch megafossils (9,709-8,954 cal. yr BP) found in peat layers in Siberia (between 70°59'N and 69°43'N) indicating that birch is a pioneer species (MacDonald et al. 2000). It should be noted, however, that the Yamal Peninsula and other regions eastward were not covered by the Fennoscandian Ice Sheet (FIS) during the Late Weichselian time period (Larsen et al. 2006). Thus, the timing of migration in the Kola Peninsula and Fennoscandia had been delayed by the retreating ice (MacDonald 2000).

Fig. 4. A 8810-yr-old birch megafossil found at the bottom of the Pousujärvi peat bog.

Pine megafossils

The subfossil pine logs (21 in number) at the Pousujärvi site were found at depths of 1-2 m below the surface of the mire. The three dated logs yielded ^{14}C-ages from 5,000±40 to 5,110±60 BP yr (5,730 to 5,900 cal. BP). On the basis of tree-ring analyses, the two oldest pine logs yielded a date of 6,006-6,054 cal. yr BP, and the three youngest 4,698, 4,385, and 4,369 cal. yr BP, respectively. Hence, pine had been present at Pousujarvi between 6,000-4,400 yr. BP. The Pousujärvi site is the very last site before Lake Kilpisjärvi on the Finnish-Norwegian border, where subfossil Scots pine logs were found. The altitude of 450 m (a.s.l.) is much lower than the nearest finding at Ailakkavaaran lompolo (68°57'N, 20°57'E at 515 m a.s.l.) by Eronen and Zetterberg (1996). Their time-envelop for the presence of pine ranges from 6,000 to 4,085 cal. yr BP, coinciding with our dates for the immigration time of pine. The stomata/pollen stratigraphy at the Toskaljavri site by Seppä et al. (2002), indicates pine immigration as early as 8,300 cal. yr BP, but this is not supported by pine megafossils.

The subfossil pine logs (37 in number) sampled from the bottom of Lake Kompsiotievanlammit yielded ages less than 3,500 yr BP as dated by the tree-ring Advance-10 master-series. Based on the radiocarbon (^{14}C) dates of subfossil Scots pine logs, pine spread through northern Fennoscandia between 8,000-7,000 yr BP and reached its maximum distribution between 6,000-4,000 cal. yr BP, followed by a gradual retreat to its modern range limit (Eronen & Hyvärinen 1987, Eronen et al. 1999, 2002, Payette et al. 2002). After the retreat of pine to its present treeline, we suspect that the tree line had been stable for 3000 yr.

Peat profile

The existence of *Equisetum* and *Menyanthes* indicate that the type of Pousujärvi mire was a wet, oligotrophic fen. The huminosity of peat is H3-4. The top layer down to 60 cm is composed of *Carex-Sphagnum* peat with gyttja-horizons containing dystrophic algae. The second layer, from 60 to 190 cm, has an abundance of *Equisetum* indicating a wet environment. However, the top of this layer is typified by *Bryales-Sphagnum* peat also contains nanoligniquids suggesting temporary drier periods. The lower part of this second layer (130-190 cm), with *Carex-Sphagnum* peat, indicates a stable fen-stage. The lowermost peat layer (190-210 cm), lying on gyttja deposits, is composed of eutrophic

Bryales-Sphagnum (H3) peat with an abundance of *Menyanthes*. The 50 cm thick gyttja layer is lying on lacustrine silt.

The sampled peat profile yielded ^{14}C-ages were consistent with the dates of pine and birch megafossils: 2,920±60 BP yr (3,060 cal. BP) at 30-cm-depth, 4,460±50 BP yr (5,050 cal. BP) at 70-cm-depth, 5,640±60 BP yr (6,410 cal. BP) at 150-cm-depth, 6140±60 BP yr (7,000 cal. BP) at 200-cm-depth, and 7,830±80 yr BP (8,600 cal. BP) at 240-cm-depth, respectively (Fig. 5). The results indicate that the maximum growth of peat was achieved between 5,050-7,000 cal. yr BP varying in rate from 0.6 mm/yr to 0.85 mm/yr. A clear climatic deterioration reduced the growth rate after 5,050 cal. yr BP, descending to 0.2 mm/yr and finally to 0.1 mm/yr. During the early Holocene, the rate was also rather low, 0.25 mm/yr. Similar to Pousujärvi pine megafossils, the timing of the maximum growth rate of the Pousujärvi peat are concurrent with the record of ^{14}C-dated pine megafossils, indicating that pine was present between 6,000-4,000 cal. yr in outermost northwest Finnish Lapland (Eronen & Hyvärinen 1987, Eronen et al. 1999, 2002). However, the maximum peat growth between 6,000-7,000 cal. yr. BP had no equivalent in the megafossils at the Pousujärvi study site.

Fig. 5. The Pousujärvi peat profile. Given numbers refer to dates (cal. BP) obtained from peat samples.

Lack of spruce

We found no macrofossil evidence to indicate the presence of Norway spruce during the Holocene in northwest Finnish Lapland. Prior to our study, a number of subfossil logs were investigated in the area, but all have been identified as pines (Eronen & Hyvärinen 1987, Eronen et al. 1999, 2002). One may argue that spruce logs were decomposed or destroyed in the lakes by wave action. However, according to our visual observations birch megafossils are present at the bottom of the small ponds, such as in the Malla Strict Nature Reserve (69°03'N, 20°38'E at 500 m a.s.l.), and birch logs were preserved through the Holocene in the peat of the Pousujärvi site. Also, fragments of early Holocene birch wood (9,730 cal. BP) beneath 3.5 m thick landslide material have been preserved (Sutinen 2005). In addition, spruce stumps were preserved in the Siberian tundra and Caledonian mountain range in Sweden through the Holocene (MacDonald et al. 2000, Kullman 2002).

On the basis of ^{14}C-dated megafossils, spruce (*Picea abies ssp. obovata*) migrated (app. 9,400 cal. yr BP) beyond its present tree limit much earlier compared to pine (app. 7,500 cal. yr BP) in Siberia (MacDonald et al. 2000). Also, spruce megafossils well above the present alpine tree limit in central Sweden indicate that spruce was part of the forest migration during the early Holocene (8,000-11,000 cal. yr BP; Kullman 2002). Thus, the conceptual migration model that advocated that spruce reached its northernmost range limit as late as 3,500 yr BP in northern Fennoscandia (Moe 1970, Kremenetski et al. 1999) appears controversial. Spruce is able to thrive at high altitudes and low temperature conditions similar to pine (Nikolov & Helmisaari 1992), hence the opposite spreading behaviour in northern Fennoscandia sounds irrational, that is spreading at the time of climatic deterioration and retreat of spruce in Siberia (MacDonald et al. 2000) and pine in Lapland (Eronen 1979, Eronen & Hyvärinen 1987), and as seen by the reduced growth of Pousujärvi peat. The Kompsiotievanlammit site lies on the present pine treeline, while the present spruce treeline is 60 km southeast of the site. If spruce arrived Lapland 3,500 cal. yr BP (Moe 1970, Kremenetski et al. 1999), why is there no evidence to indicate a spruce migration with a similar topoclimatic position as pine. Also, the presence of Norway spruce after 4,000 cal. yr BP in the Toskaljavri pollen stratigraphy (Seppä et al. 2002) appears vague. If spruce migrated to the Kilpisjärvi area, it should have left macro-signatures in the peat layers of Pousujärvi and other investigated peat sites in the region (Eronen & Zetterberg 1996).

With respect to the present spruce timberline in the transition between the mafic Tanaelv belt and felsic Lapland granulite belt in Finnish Lapland (Sutinen et al. 2005), we argue that in contrast to Siberian cases (MacDonald et al. 2000), spruce has not been a ubiquitous member of the Holocene tree succession sequence in northern Fennoscandia. Since spruce and pine have distinctly different site requirements with respect to soil water availability and nutrients (Sutinen et al. 2002), the reverse behaviour between the timberlines and migration patterns of Eurasian spruce and pine species may be associated with differences between the soils derived from the rocks of the Baltic Shield and those derived from Cambrian sedimentary rocks in subarctic Siberia.

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Geological Survey of Finland, Special Paper 46, 99–102, 2007.

POLLEN DIAGRAM OF THE LATE PLEISTOCENE INTERSTADIAL DEPOSITS FROM THE KOLA PENINSULA, RUSSIA

by
Alla Sharapova and Ljudmila Semenova[1]

Sharapova, A. & Semenova, L. 2007. Pollen diagram of the Late Pleistocene interstadial deposits from the Kola Peninsula, Russia. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 99–102, 2 figures.

A 40-meter long sediment succession from the Svjatoy Nos Cape, Murman Coast of the Kola Peninsula, has been analyzed. Lithostratigraphic investigations combined with pollen analysis have revealed new evidence on the Late Pleistocene history of the area. Five sedimentary units have been identified. These units represent various sedimentary environments including coastal marine sediments (the oldest unit I), marine sediments (unit II), glaciofluvial deposits (unit III), till (unit IV) and soil (unit V). Marine deposits (unit II) yielded abundant assemblages of pollen, belonging to 22 morphotypes. The pollen percentage diagram was divided into two pollen assemblage zones. The zones are characterized by dominance of *Betula sect. Albae and Ericales*. The upper pollen zone is marked by rapidly increasing values of *Betula sect. Nanae, Cyperaceae* and *Artemisia*. The pollen record is interpreted to reflect the Late Pleistocene interstadial conditions with birch-forest tundra and tundra vegetation.

Key words (GeoRef Thesaurus, AGI): paleobotany, sediments, biostratigraphy, pollen analysis, interstadial environment, Pleistocene, Kola Peninsula, Russian Federation.

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INTRODUCTION

The biostratigraphy of the Late Quaternary sequence from the Kola Peninsula is poorly known. There are a few palynological studies of the Pleistocene deposits from this area (Nikonov & Vostruhina 1964, Armand 1969, Armand et al. 1969, Evserov et al. 1980). The aim of the present study is to establish a palynostratigraphy and vegetation history of the Late Pleistocene interstadial event from the northeastern part of the Kola Peninsula.

STUDY AREA

The Kola Peninsula is situated in northeast Fennoscandia and is bounded by the Barents Sea to the north and the White Sea to the south. The climate is influenced by the relatively warm Murman Coast current, a branch of the North Atlantic current. Mean January temperatures range from -5 to -9°C for the Barents shoreline, and from -10 to -14°C for the central part of the Kola Peninsula. Mean July temperatures range from 8 to 10°C and from 11 to 14°C respectively (Koshechkin et al. 1975). Study site SN-1 is located in the northeastern part of the Kola Peninsula on the Barents Sea coast (Fig. 1). The vegetation of the study area belongs to tundra zone (Ramenskaya 1983).

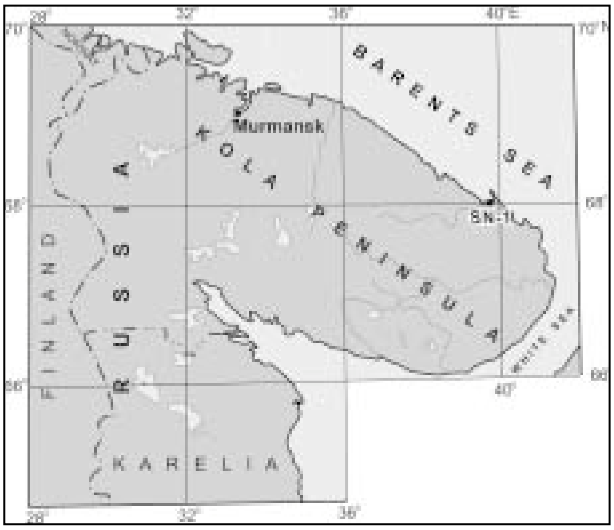

Fig. 1. Map of the study area.

METHODS

Pollen analysis was carried out on samples with a vertical thickness of 1 cm. Wet weight for the samples was 50 g. In general, treatment followed the method described by V. Grychuk (1940), including disintegration in 10% KOH and separation in heavy liquid KI + CdI_2, density 2.3. For each sample, 300 arboreal pollen grains were counted on magnification x 600; nonarboreal pollen and spores were also tallied. Plant nomenclature follows Vascular Plants of Russia and Adjacent States (1995). Pollen diagrams were made using the TILIA program (Grimm 1990). Percentages of pollen and spore taxa were calculated per total sum of pollen and spores.

RESULT AND DISCUSSION

A 40-meter long sediment succession from the Svjatoy Nos Cape, Murman Coast of the Kola Peninsula was analyzed. Five sedimentary units were identified. The lowermost lithological unit I, coastal marine deposits (37.6-32.6 m), overlies the crystalline bedrock and consists of sand, gravel and boulder. Lithological unit II, marine deposits (32.6-5.3 m), is composed of clay. Lithological unit III, glaciofluvial deposits (5.3-2.8 m) consists of sand, gravel and boulder. Lithological unit IV, glacial deposits (2.8-0.7 m), represents till. Lithological unit V (0.7-0 m) consists of soil (Fig. 2). Marine deposits (unit II) were studied for spore and pollen analysis. Twenty-two morphotypes of Quaternary pollen and spores were identified. The pollen percentage diagram was divided into two pollen assemblage zones (Fig. 2). Trees and shrubs dominated in the pollen assemblages of zone 1 (subzone 1a, 1b, 1c, 1d; 30.6-10.2 m) and zone 2 (10.2-5.3 m), changing from 47 to 70%. The values of spores were 22-48%; contents of herbs were 5-18%. These zones are characterized by the dominance of *Betula sect. Albae* and *Ericales*. Pollen zone 2 is marked by rapidly increasing values of *Betula sect. Nanae, Cyperaceae* and *Artemisia sp*.

Pollen zone 1 reflects birch-forest tundra vegetation, and pollen zone 2 marks tundra vegetation. The surface sediment sample from the Holocene soil (unit V) is typified by the dominance of *Pinus sylvestris* (48%) and *Betula sect. Albae* (28%). The conclusion is that low contents of *Pinus sylvestris* in pollen zones 1 and 2 (5-10%) indicate cooler climatic conditions in comparison with present day. The pollen record is thus interpreted to represent the Late Pleistocene interstadial event.

The pollen assemblages of zone 1 and zone 2 are similar to the pollen records from the Middle Valdaian (Middle Weichselian) deposits from the south-western part of the Murmansk region (Evserov et al. 1980) and from the Peräpohjola region, northern Finland (Korpela 1969).

Radiocarbon data from the studied site obtained previously (46,540±1,770 ^{14}C BP (Gudina & Evserov 1973); 41,900±1,290 ^{14}C BP, 43,160±1,330

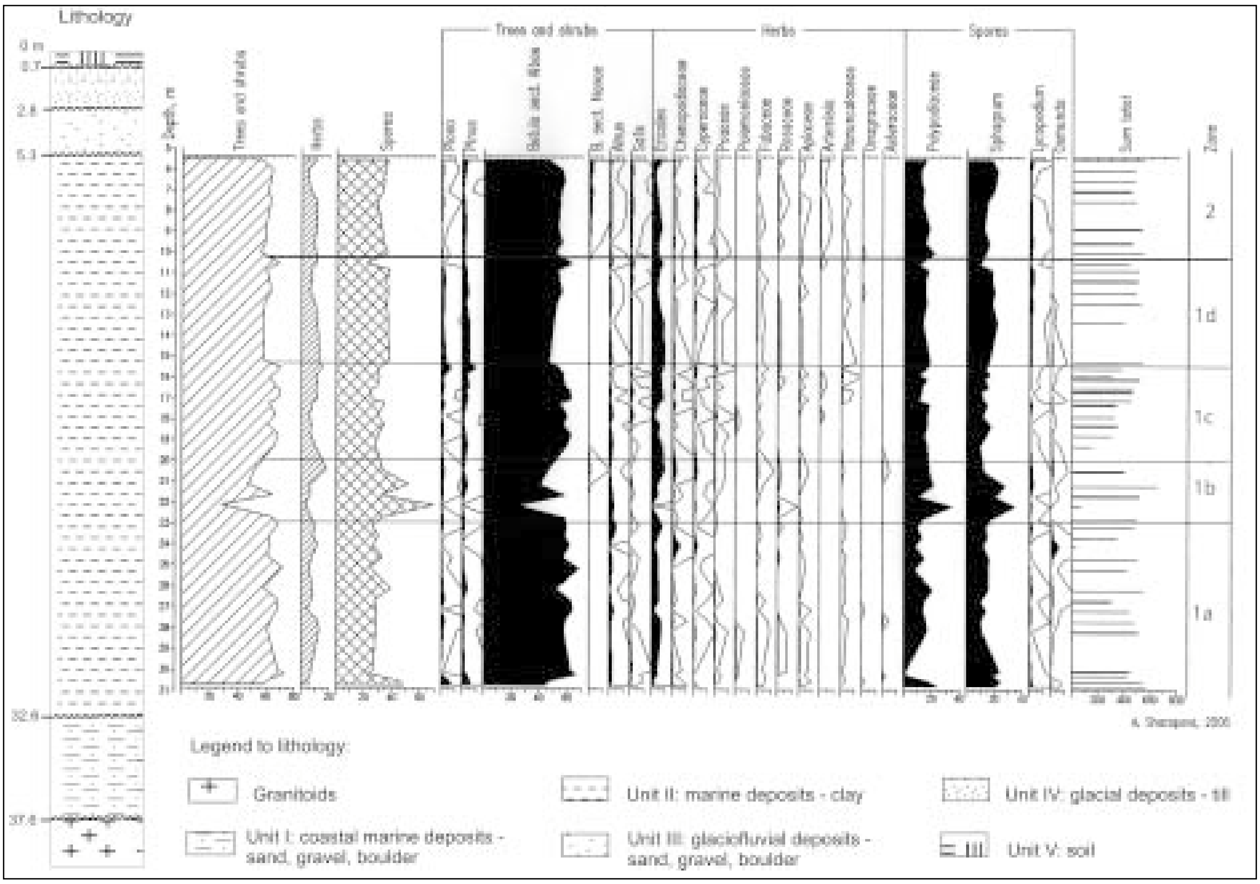

Fig. 2. Pollen percentage diagram of the Late Pleistocene deposits from the Svjatoy Nos Cape.

^{14}C BP, 47,940±1,820 ^{14}C BP (Arslanov et al. 1981)) point to the Middle Valdaian (Middle Weichselian) age of marine deposits. But Th-230/U-234 date – 97,000±4,000 BP (Arslanov et al. 1981) does not confirm this conclusion. Further investigation will clarify the chronostratigraphic position of these deposits.

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POLLEN AND IR-OSL EVIDENCE FOR PALAEOENVIRONMENTAL CHANGES BETWEEN CA 39 KYR TO CA 33 KYR BP RECORDED IN THE VOKA KEY SECTION, NE ESTONIA

by
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Bolikhovskaya, N. S. & Molodkov, A. N. 2007. Pollen and IR-OSL evidence for palaeoenvironmental changes between ca 39 kyr to ca 33 kyr BP recorded in the Voka key section, NE Estonia. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 103–112, 4 figures.

The sediment sequences of the Voka site situated on the south-eastern coast of the Gulf of Finland were analysed to reconstruct the environmental evolution of northern Estonia over the last Ice Age. Our reconstruction is based on an optically-dated record of pollen-induced palaeoclimate (palaeoenvironmental) proxies. Plotted against depth, palynological analysis and luminescence datings of the samples produced a dated sequence ranging from ca 39 to ca 33 kyr BP, which reveal the specific features of the palaeoenvironmental variations before the last glacial maximum (LGM). Representative pollen spectra from 45 samples provided convincing evidence of noticeable changes in vegetation and climate in northeastern Estonia. Two intervals of severe climate and two considerably milder ones were recognised. Matching of the Voka chronoclimatic pattern to the Greenland ice core variations shows a good fit with the palaeoclimatic signals recorded in Greenland ice cores between Greenland Interstadials 8 and 5.

Key words (GeoRef Thesaurus, AGI): palaeoenvironment, palaeoclimatology, palaeobotany, sediments, pollen analysis, geochronology, optically stimulated luminescence, Pleistocene, Weichselian, Voka, Estonia.

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INTRODUCTION

The last Ice Age is generally assumed to have been characterised by pronounced climatic variations, recorded in many geological archives around the Baltic sphere (Mangerud 1981, Grichuk 1982, Lagerback & Robertsson 1988, Behre 1989, Mamakowa 1989, Liivrand 1991, Donner 1995, Kondratienė 1996, Helmens et al. 2000, Kalnina 2001). However, the timing and reconstruction of palaeoenvironmental evolution on the basis of available fragments of geological record remains a problem in Quaternary geology. There are often gaps in the geological record, and the absolute age of its constituents is frequently unknown. Local stratigraphic schemes are often difficult to correlate with interregional time divisions and with the global palaeoclimatic levels of the ice-core and deep-sea oxygen isotope records. This especially holds true in the Baltic countries where the late Pleistocene glacial erosion over this area could have been very intensive and, therefore, palaeoenvironmental records here may consist of a number of hiatuses. As a result, even the number of Järva (=Weichselian) glaciations or big stadials in Estonia, the northernmost of the three Baltic states, is not yet precisely known (Raukas 2003). In the Estonian Weichselian, two till beds (Valgjärv and Võrtsjärv subformations) are distinguished. The till beds correspond to two Järva glaciation cycles, each of which lasted some 50–55,000 years (Raukas et al. 2004); however, this arouses doubts, see, for example, Kalm (2006). As well, it is not yet clear whether the oldest Valgjärv till belongs to the lower or middle Weichselian (Raukas 2003). The till is widely distributed in central and southern Estonia but is absent from the bedrock areas and other regions of intense glacial erosion in the northern and north western parts of the country (Raukas et al. 2004). For the corresponding early Weichselian glaciation, Kalm (2006) leaves only some 25 kyr (68–43 kyr BP). Therefore, it becomes understandable why different, often contradicting schemes of stratigraphic subdivision and palaeoenvironmental interpretations of past climates and sedimentary history of the region during the last Ice Age have been proposed in the last decades (Kajak et al. 1976, Raukas & Kajak 1995, Gaigalas 1995, Satkūnas 1997).

The survival of well-preserved, in situ pre-LGM deposits is very rare in northern Estonia. In this area, glacial erosion prevailed and the thickness of the Quaternary deposits is usually less than 5 m. Therefore, interdisciplinary study of those sections that do exist, such as Voka site, can potentially provide a unique opportunity to reconstruct the palaeoenvironmental history of the region in greater detail.

The aim of this study was to identify the main palaeoenvironmental stages in northern Estonia using palaeoclimate proxies based on detailed palynological analysis of the deposits from the new late Pleistocene section at the Voka site, north eastern Estonia. A depth-age model in this study was constructed using regression analysis of the infrared optically stimulated luminescence (IR-OSL) dating results obtained on the samples collected from this section (Molodkov 2007).

STUDY SITE AND RESEARCH METHODS

The Voka site is located on the south eastern coast of the Gulf of Finland where the Estonian part of the Baltic Clint is interrupted by an about 2.2-km-wide depression – the Voka Clint Bay (Figs. 1A, B). The latter, in its turn, is intersected by a deep canyon-like palaeoincision. According to drilling data, most of the incision is filled with fine-grained sand and silt with a total thickness of at least 147 m. Layers of gravely sand, coarse- to medium-grained sand and clays also occur. The northernmost part of the valley deepens very abruptly, with the bottom gradient reaching 12 m per km and the slope gradient is more than 500 m per km (Molodkov et al. 2007).

The deposits studied are exposed along the Gulf coast in the V3 outcrop (59°24.86´N, 27°35.88´E) situated about 500 m to the east from the mouth of the Voka (Vasavere) River (see Fig. 1B). Immediately on the shore, the height of the outcrop is about 22 m a.s.l. The sandy to clayey sedimentary sequence comprises at least two lithostratigraphic units – A and B (Fig. 2). Topwards unit A in the V3 outcrop is overlain by about 1.1-m-thick brownish-grey to light-brown soil resting upon ca 0.7-m-thick layer of light-brown silt-like deposits showing signs

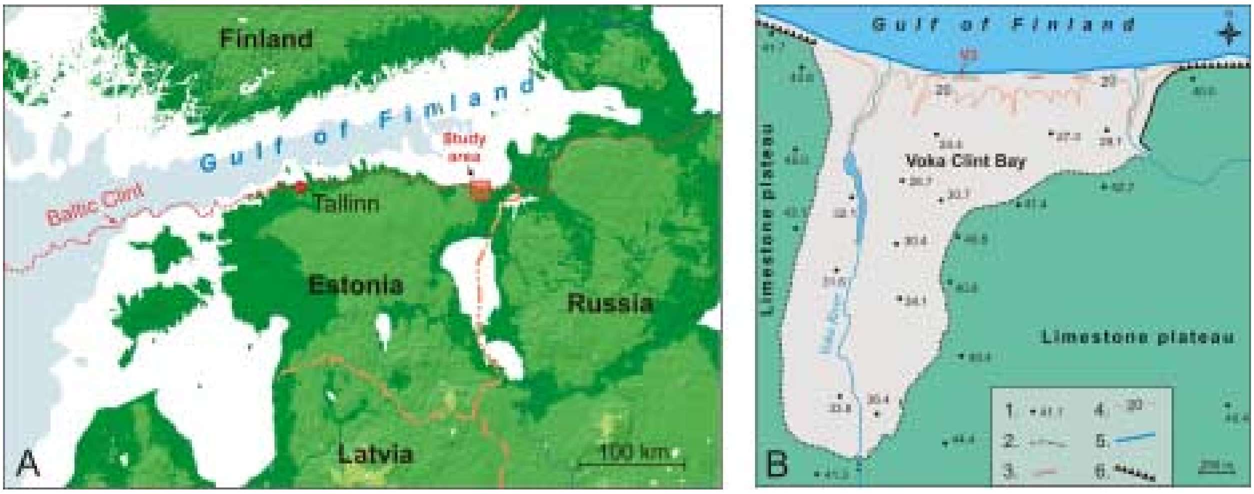


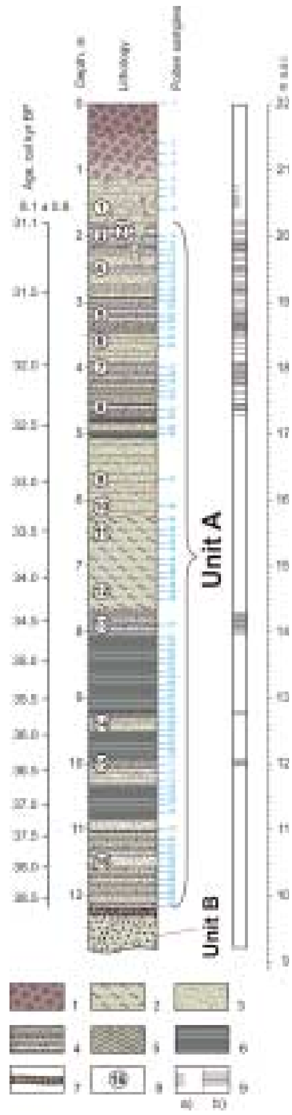
Fig. 1. Location map of the study area (A) and the scheme of the Voka Clint Bay with locations of the outcrop V3 (B); 1– altitudes above sea level in metres; 2 – Clint Bay boundary; 3 – outcrop; 4 – 20 m contour line; 5 – coastline; 6 – clint.

of soil formation, densely penetrated by roots and filled krotovina-like irregular tunnels made probably by burrowing animals. The boundary between units A and B is marked by a gravel bed. The thickness of unit A in the V3 section is about 10 m. unit A in this section is characterised by alternating laminated fine sand and clay with dispersed particles of lower Ordovician *Dictyonema* argillite. Unit B is represented by coarse-grained cross-bedded sands. The maximum visible thickness of the unit (shown only partly in Fig. 2) is about 15 m.

More than 90 samples were taken from unit A of the V3 section and from overlying soil deposits for detailed palynological analysis. In the present study, the palyno-climatostratigraphic method was applied to identify pollen zones in the record. The standard palynological technique, which includes treatment by 10% HCl, 10% KOH and acetolysis solution, was applied for extraction of pollen and spore grains from deposits (Grichuk & Zaklinskaya 1948). Samples of sands at the depths between 1.8–5.0 m turned out to be poor in pollen and spores.

Since the deposits were devoid of macroscopic organic remains, the infrared optically-stimulated luminescence (IR-OSL) technique was applied to produce an absolute chronology of the deposits. Fifteen samples from unit A of the sections were measured with IR-OSL and an age model was developed (Molodkov 2007) to provide a time scale for the unit and a temporal base for palaeoenvironmental reconstructions (all IR-OSL ages reported in this paper were obtained in the Research Laboratory for Quaternary Geochronology, Tallinn; the IR-OSL ages given here are in calendar years before present, BP; the dating results obtained for unit A from the

Fig. 2. Stratigraphic column of unit A of the V3 section at the Voka site. The numbered circles show the location of the IR-OSL sampling points. Shown to the right of the columns are the positions of pollen sampling points and the variations with depth of the *Dictyonema* argillite distribution in the deposits along the sections. Calendar time scale for unit A of the V3 section is indicated on the left (for detail see Molodkov 2007); 1 – soil; 2 – silt; 3 – sand; 4 – horizontally bedded sands; 5 – crossbedded sands; 6 – laminated clayey deposits; 7 – marker bed; 8 – IR-OSL sampling points; 9 – interrupted (a) and varve-like (b) layers of Dictyonema-bearing sediments.



exposure V3 are available in Molodkov (2007) on Quaternary Geochronology Online).

Such an integrated approach based on two independent methods applied to one and the same source of climate-stratigraphic information gives a unique opportunity to reveal specific features of pre-LGM palaeoenvironmental variations.

RESULTS OF PALAEOENVIRONMENTAL RECONSTRUCTIONS

Plotted against depth, palynological analysis and luminescence datings of the samples taken from the section produced a dated sedimentary sequence within the whole unit A (between the depths of 12.15 m and 1.80 m) covering the period from ca 39 to 31 kyr BP (Fig. 2). Deposits from the depths between 12.15 m and 5.00 m reveal specific features of pre-LGM palaeoenvironmental variations during the time span of ca 6 kyr.

The deposits overlying unit A are most likely post-LGM in age as indicated by an IR-OSL date of 8.1 ± 0.8 kyr for the sample taken from these deposits at a depth of 1.60 m (Molodkov 2007). According to several preliminary age determinations (ibid), the dated deposits of unit B span the period from at least 115 kyr to at least 70 kyr. The presence of older deposits (>115 kyr) is expected in the lower part of the sections currently unavailable due to the thick talus at the base of the outcrops. Representative pollen spectra derived from 45 samples taken at depths between 12.15 m and 5.00 m provide convincing evidence of noticeable changes in vegetation and climate in northern Estonia during the period from ca 39 to 33 kyr BP.

The results of palynological analysis of the studied sediments are presented on the spore-pollen diagram (Fig. 3). Most of the taxa of the studied palynoflora are grouped by genera and families to

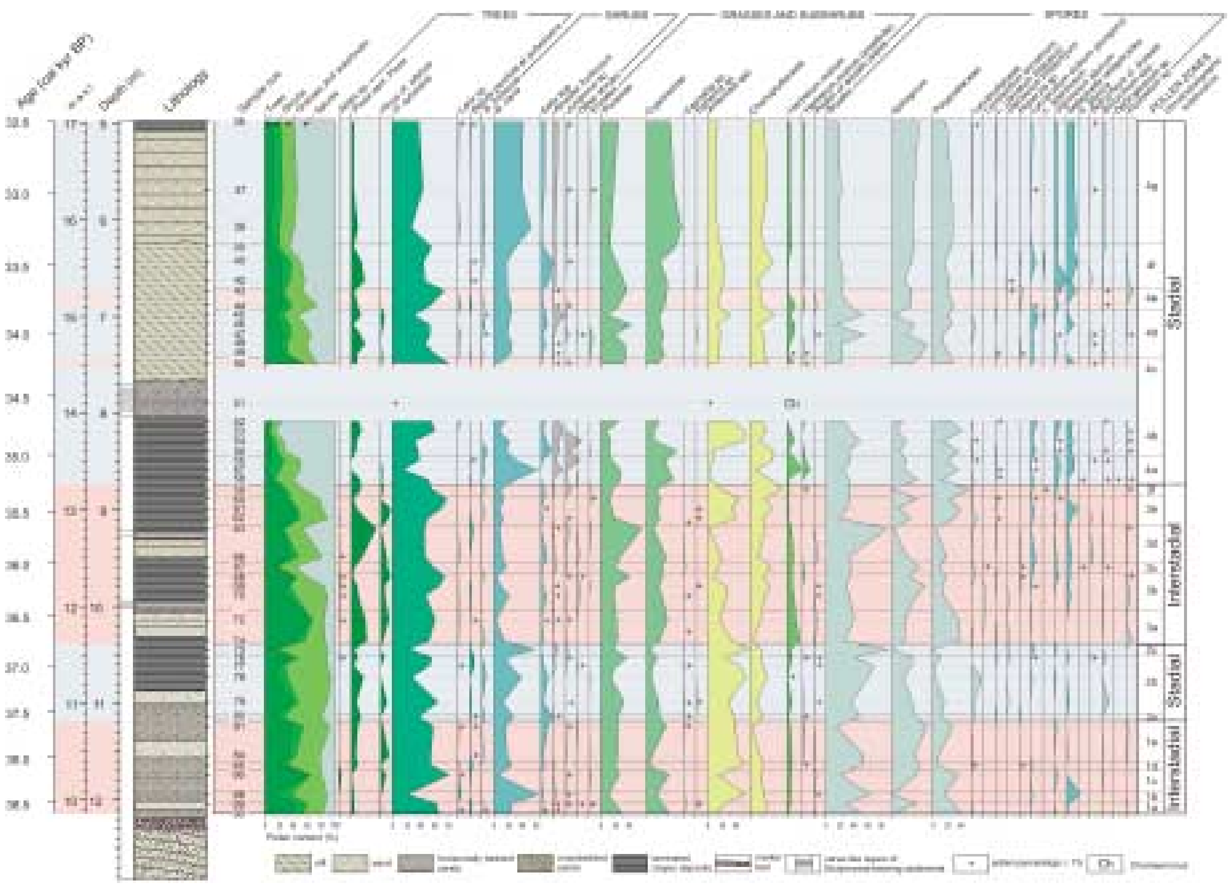


Fig. 3. Pollen and spore percentage diagram from unit A of the V3 section (5.00–12.15 m).

make the diagram more compact. The palynological data obtained testify that the period of accumulation of the middle Weichselian (=middle Valdai, middle Järva) sediments under consideration corresponds to four climatostratigraphic subdivisions – two intervals of severe climate and two interstadials with remarkably milder climatic conditions.

The pollen data from the middle Weichselian deposits are rather scarce in Estonia. Pollen assemblages recovered from the periglacial middle Weichselian deposits in Tõravere, Valguta, Savala and other sections were interpreted by E. Liivrand (1991, 1999) as evidence of cold stadial environments with widely spread tundra and xerophytic communities (with dominance of dwarf birch *Betula nana*, wormwood *Artemisia*, wormseed *Chenopodiaceae, Poaceae/Gramineae* and *Cyperaceae*). No middle Weichselian interstadial warmings have been identified up to now. In this connection, we shall consider in more detail the climatostratigraphic subdivision of the investigated deposits as well as the changes that took place in vegetation and climate during the period of sedimentation.

The 1st Voka interstadial (VIS-1A, 38.6 – 37.6 kyr BP)

The first period of relative warming (38.6–37.6 kyr BP) is revealed by pollen assemblages between 11.20–12.20 m (Fig. 3). At the Voka site, this interstadial was marked by the dominance of mainly periglacial forest-tundra, locally with open forest patches of spruce (*Picea sect. Picea*) and common pine (*Pinus sylvestris*), with a mixture of larch *Larix* and Siberian cedar pine *Pinus sibirica*.

Five pollen zones (1a–1e) corresponding to five phases of vegetation and climate evolution reflect dynamics of palaeoenvironmental changes during this interval. Periglacial forest-tundra comprising patches of pine forests (with an admixture of *Larix sp.*) prevailed during the first phase (pollen zone 1a; 12.0–12.2 m). Later, in the second phase (pollen zone 1b; 11.9–12.0 m), the climate became colder and dryer. This resulted in an increased abundance of cryophytes and xerophytes in the vegetation – *Betula sect. Nanae, Alnaster fruticosus, Selaginella selaginoides, Artemisia* (including *Artemisia subgenus Seriphidium*), *Chenopodiaceae*, etc. Patches of larch open forests occasionally occurred. Open tundra-steppe landscapes dominated.

The most significant warming within this interstadial and expansion of coniferous trees over the study area took place during the third phase (pollen zone 1c; 11.7–11.9 m). *Pinus sylvestris* accounted for more than 90% in the relevant palynospectra. Pine forests and open forests of common pine (*Pinus sylvestris*) dominated. The fourth phase (pollen zone 1d; 11.6–11.7 m) is characterised by the dominance of periglacial forest-tundra with spruce-pine forests, in combination with yernik (formed by the *Betula nana* and *Alnaster fruticosus*), and *Chenopodiaceae-Artemisia* communities.

During the fifth phase (pollen zone 1e; 11.2–11.6 m), the studied area was occupied by periglacial forest-tundra with Siberian cedar pine-spruce-common pine forests with a mixture of larch, along with steppe and shrubby (*Betula nana* and *Salix sp., Alnaster fruticosus*) communities. The steppe biotopes were prevailed *(Poaceae, Asteraceae, Primulaceae, Onagraceae, Primulaceae, Polygonaceae, Rumex sp., Urtica sp., Plantago media, Ranunculus sp*., etc.) and wormwood (*Artemisia subgenus Euartemisia, A. subgenus Seriphidium*) associations.

The 1st Voka stadial (VS-1A, 37.6 – 36.8 kyr BP)

The first period of warming (VIS-1A) was followed by a relatively short interval of a colder and dryer climate, which lasted some 800 years (37.6–36.8 kyr BP; 10.4–11.2 m). According to palynological data (see Fig. 3), the prevailing landscapes were tundra-steppe and tundra-forest-steppes.

The herb-dwarf shrub pollen (mainly *Artemisia, Chenopodiaceae, Poaceae, Cyperaceae*) and spores (*Bryales, Sphagnum*) predominate in all pollen assemblages (NAP – 41 to 70%; Spores – 6 to 20%). Besides, the pollen spectra are characterized by high percentages of cryophytes and xerophytes, such as *Betula nana, Alnaster fruticosus, Artemisia (including A. subgen. Seriphidium, A. subgen. Dracunculus), Chenopodiaceae, Ephedra, Lycopodium dubium, Diphazium alpinum, Selaginella selaginoides*.

Pollen zones 2a, 2b and 2c correspond to three phases of environmental evolution and reveal transformations of vegetation during this stadial episode.

In the first phase (pollen zone 2a; 11.1–11.2 m) open tundra-steppe environments predominated.

In the second phase (pollen zone 2b; 10.5–11.1 m) humidity increased, and the tundra-forest-steppes with patches of Siberian cedar pine-spruce-common pine light forests became dominant.

In the third phase (pollen zone 2c; 10.4–10.5 m), the climate turned drier and tundra-steppes prevailed again.

During the first and third phases, tundra-steppe vegetation was most widespread in the surroundings of Voka. The distinctive features of the long-term second phase were the more humid climate, expansion of patches of coniferous (Siberian cedar pine–spruce–common pine) light forests and predominance of tundra-forest-steppes.

The 2nd Voka interstadial (VIS-2A; 36.8 – 35.3 kyr BP)

The next interstadial (36.8–35.3 kyr BP; 8.75–10.40 m), much warmer than the previous one, was characterised by a new expansion of boreal forest species onto the area under study. Six climatic-phytocoenotic phases were reconstructed, reflecting changes in the plant communities during this interval. These successions show that at least twice – during wetter and warmer climatic phases – periglacial forest-tundra (locally with open spruce and pine forests) was replaced by more humid landscapes with prevailing northern taiga communities.

During the first phase, the Siberian cedar pine–spruce–common pine (*Pinus sylvestris*) forests and parkland forests dominated, corresponding to pollen zone 3a (10.1–10.4 m). In the evolution of vegetation, this was the warmest interval of the whole interstadial. In the dendroflora, the total boreal pollen accounted for about 70–90%. The second phase (pollen zone 3b; 9.65–10.1 m) was prevailed by periglacial forest-tundra with patches of Siberian cedar pine–pine open forests. In approximately the middle of this interstadial – in the third phase (pollen zone 3c; 9.55–9.65 m) – a short interval with a significantly colder climate and domination of periglacial tundra is recorded.

During the forth phase (pollen zone 3d; 9.15–9.55 m) periglacial forest-tundras with patches of pine-spruce and Siberian pine-spruce-pine light forests dominated.

Periglacial light forests of Siberian cedar pine (*Pinus sibirica*), spruce (*Picea sect. Picea*) and common pine (*Pinus sylvestris*) prevailed in the Voka area during the fifth phase characterized by pollen zone 3e (8.85–9.15 m), reflecting a new warming within VIS-2A. During the latest sixth phase (pollen zone 3f; 8.75–8.85 m), periglacial forest-tundras with small patches of spruce–pine open woodlands predominated in the studied area.

The 2nd Voka stadial (VS-2A; 35.3 – 32.6 kyr BP)

During the last interval (VS-2A; 35.3–32.6 kyr BP), periglacial tundra remained the dominant landscape in the region. The considerably colder climate and expansion of permafrost resulted in the appearance of wetlands of moss and grass types. The pollen assemblages were dominated by *Betula nana*, *Poaceae* and *Cyperaceae*, together with the spores of *Sphagnum, Bryales, Selaginella selaginoides, Lycopodium dubium, Diphazium alpinum* and other typical tundra species. This suggests that the interval was close to the final part of the middle Weichselian preceding the late Weichselian ice sheet expansion into the region.

Pollen-and-spore spectra recovered from the deposits at the depths between 8.75 and 4.95 m allowed for reconstruction below the presented seven successive phases in the evolution of vegetation and climate during this second cold episode.

In the first phase (pollen zone 4a; 8.45–8.75 m), periglacial tundra dominated with a prevalence of yernik bushes (from dwarfish birch *Betula nana*) and shrubby communities (from juniper *Juniperus sp*., willow *Salix sp., Alnaster fruticosus*, etc.) and grass-moss bogs.

In the second phase (pollen zone 4b; 8.05–8.45 m), a distinctive feature of vegetation was the almost ubiquitous expansion of periglacial tundra with the domination of moss bogs and the presence of small patches of tundra shrubs.

Sands in the depth interval 7.55–8.05 m contained only single pollen grains.

In the third phase (pollen zone 4c; 7.45–7.55 m), periglacial forest-tundra with small patches of pine light forests (of *Pinus sylvestris*) occupied the Voka area under warmer climate conditions.

In the fourth phase (pollen zone 4d; 6.95–7.45 m), the forested areas were considerably reduced due to the cooling of the climate. The role of microthermic tundra plants (*Betula nana, Alnaster fruticosus, Selaginella selaginoides, Lycopodium dubium, L.appressum, Diphazium alpinum*, etc.) increased. Periglacial tundras dominated. *Pinus sylvestris* and *Larix sp*. were edificators in forest stands of rare forest-tundra communities.

A new short warming led to reduction of tundra shrubby coenoses in the fifth phase (pollen zone 4e; 6.70–6.95 m). The significance of periglacial forest-tundras and boreal tree species (*Pinus sylvestris*, *Larix*, etc.) increased.

In the sixth phase (pollen zone 4f; 6.25–6.70 m), the incipient climate cooling triggered a new expansion of grassy cryophytes (*Selaginella selaginoides, Diphazium alpinum, Lycopodium dubium, L. appressum*, etc.) and domination of periglacial tundras.

In the final seventh phase of this stadial (pollen zone 4g; 4.95–6.25 m), periglacial tundra predominated in the Voka area with a prevalence of yernik bushes (from dwarf birch *Betula nana*) and grass-moss bogs.

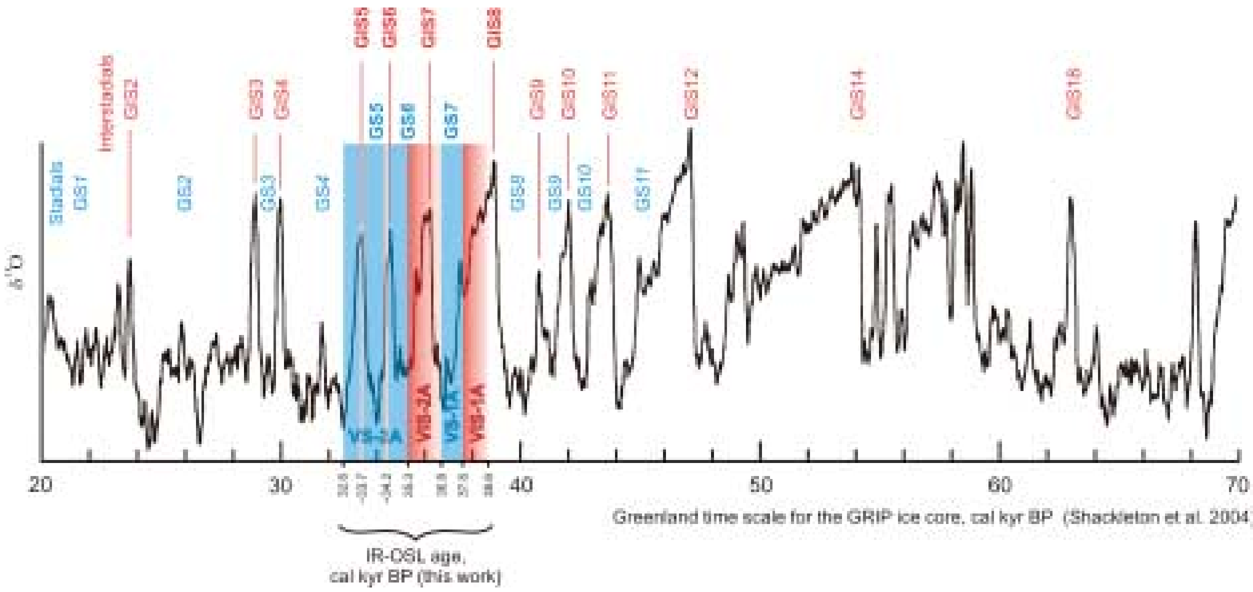


Fig. 4. Matching Voka chronoclimatic pattern (this work) to the Greenland ice-core variations (Shackleton et al. 2004). The best fit lies between Greenland Interstadial 8 (GIS8) and GIS5.

Thus, two minor warmings are recognised within the second cold event (VS-2A) at depths of about 7.50 m and 6.70–6.95 m corresponding to the ages of 34.2 kyr and between 33.7 and 33.8 kyr BP, respectively. During these short-term warmings, the dominating periglacial tundras were replaced by periglacial forest-tundras with small patches of light forests, in which Pinus sylvestris served as an edificator.

DISCUSSION

The reconstructed paleoenvironmental changes give evidence for specific individual features of the mid-Weichselian palaeoclimatic stages.

The obtained pollen data show that the share of dark coniferous trees – spruce and Siberian cedar pine – was considerable in the forest communities during the interstadial between ca 38.6 – 37.6 kyr BP, whereas during the younger interstadial ca 36.8 – 35.3 kyr BP spruce (*Picea sect. Picea*) and *Pinus sibirica* were far less abundant. Forest biotopes of this warmer interstadial were rather dominated by larch (*Larix sp.*).

Palaeoenvironments of each cold interval also had their own characteristic features. This primarily concerns the dominant zonal types of vegetation. Climatic conditions of the oldest interval were more arid, and the study territory was covered by tundra-steppe and tundra-forest-steppe. The open grass-low bush biotopes were dominated by *Chenopodiaceae-Artemisia* associations. During a long cold epoch that lasted for about 2500 years from ca 35.3 kyr BP till at least 32.6 kyr BP, the Voka area was characterized by periglacial tundra with grasses and sedges dominating the open landscapes. Also, such arctic alpine and hypoarctic tundra grass species as *Selaginella selaginoides*, *Diphazium alpinum, Lycopodium dubium, L. appressum, Botrychium boreale* were abundant.

The data on the duration of the reconstructed interstadials and stadials during the Weichselian glacial epoch are of critical importance for detailed chrono- and climatostratigraphy. Based on the obtained IR-OSL data, the duration of the established interstadials is estimated at approximately 1000 and 1500 years. Due to the incompleteness of the geological record of the most ancient of the interstadials, caused by a relatively long break in sedimentation, its duration was not less than 1500 years. The duration of the reconstructed mid-Weichselian cold stadials is even more different. The first cooling lasted for about 800 years (ca 37.6 – 36.8 kyr BP), but the next one was at least 2700 years long (ca 35.3 – 32.6 kyr BP). These data together with the obtained floristic, phytocoenotic, and climatostratigraphic characteristics of the investigated interstadials and stadials should be considered when identifying and correlating palaeogeographical events of the Weichselian (Valdai, Järva) glacial epoch. Our data unambiguously demonstrate that the studied sediment sequence in the Voka section is by no means a late glacial formation. This is substantiated by comparison of the climatophytocoenotic features and chronological extent of the studied intervals with analogous data on the late glacial events in Estonia and adjacent regions.

Firstly, the late glacial warm and cold intervals had a shorter duration than the stratigraphical units investigated in the Voka section. Secondly, the Voka interstadials strongly differ from the Allerød, which is well represented in the Baltic countries, also in the dominant type of palaeovegetation. Forests of the reconstructed warm epochs were lacking birch stands and were dominated by coniferous communities. In southern and northern Estonia, birch was predominate among arboreal species during the first half of the Allerød, and only during its second half was replaced by pine (*Pinus sylvestris*) (Pirrus 1969, Pirrus & Raukas 1996, Kabailienė & Raukas 1987). During the Allerød, forests composed of *Betula sect. Albae* and alder (*Alnus sp.*) were widespread to the east from Estonia, in the vicinity of the Lake Il'men' (Razina & Savel'eva 2001). Further north, in the southern Karelia, the birch Betula sect. Albae, including *Betula pubescens*, was also a dominant species (Elina et al. 1995, Wohlfarth et al. 2002).

Matching our Voka chronoclimatic pattern to the Greenland ice-core variations shows that there is only one good fit with the palaeoclimatic signals recorded in Greenland ice cores, which exists between Greenland Interstadial 8 (GIS8) and GIS5 (Fig. 4). The first Voka interstadial VIS-1A identified within unit A corresponds to GIS8. The second Voka warm event VIS-2A is clearly comparable with GIS7. It is highly probable that two warm spells identified in the Voka section during the second Voka stadial VS-2A can be correlated with GIS6 and GIS5. As well, this Voka stadial VS-2A can be correlated with Greenland Stadial 6 (GS6), GS5 and with the initial part of GS4.

As in the Greenland ice core record, the second interstadial in the Voka section is warmer than the previous one. It also seems that the initial part of the interstadial correlative to GIS8 is absent in the Voka section because the corresponding part of the sediments is interrupted by a gravel bed at the bottom of unit A. As well, like in the ice core, the second Voka stadial VS-2A is also somewhat colder than the previous one. Thus, it seems that there are quite a few coincidences between features in the ice core and the Voka records. In view of this, there is a possibility that these coincidences are not accidental.

CONCLUSIONS

It was established in the present study that optical ages and palynological features derived from unit A of the V3 section mark the time of the existence of relatively complex and changeable environmental conditions during the middle Weichselian pleniglacial.

Representative palynological data obtained at the Voka site have provided convincing evidence of the occurrence of two severe and two considerably milder climate intervals in the time span ca 39–33 kyr.

The first interstadial at the Voka site (38.6–37.6 kyr) was marked by the dominance of mainly periglacial forest-tundra, locally with open forest patches of spruce and common pine with a mixture of larch and Siberian cedar pine. During the following relatively short (37.6–36.8 kyr) interval with a colder and dryer climate, landscapes of tundra-steppe and tundra-forest-steppes prevailed. The next interstadial (36.8–35.3 kyr), much warmer than the previous one, was characterised by a new expansion of boreal forest species into the study area. The last interval (35.3–32.6 kyr) was characterised by landscapes where a periglacial tundra type of vegetation dominated. The considerably colder climate at that time and expansion of permafrost resulted in the appearance of wetlands of moss and grass types. This suggests that this interval was close to the final part of the middle Weichselian preceding the late Weichselian ice sheet expansion into the region.

The Voka outcrops studied do not display the assumed late-glacial deposits. Instead, it was established that the deposits exposed in the Voka outcrops demonstrate a general similarity between the Voka and Greenland ice-core records. Matching the Voka chronoclimatic pattern to the Greenland ice-core variations shows a good fit with the palaeoclimatic signals recorded in Greenland ice cores between Greenland Interstadials 8 and 5.

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DATED – A GIS-BASED RECONSTRUCTION AND DATING DATABASE OF THE EURASIAN DEGLACIATION

by
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Gyllencreutz, R., Mangerud, J., Svendsen, J.-I. & Lohne, Ø. 2007. DATED – A GIS-based Reconstruction and dating Database of the Eurasian Deglaciation. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 113–120, 3 figures.

The goal of the ongoing project DATED is to describe and document the ice growth towards the LGM and the deglaciation of the large ice sheets in northwest Eurasia. Digitized ice margins and other relevant published features are compiled in a geographical information system, which is coupled to a database with dates (^{14}C, OSL, TL, cosmogenic exposure, clay-varve). The first version of the database and the GIS will be available on the internet in 2007, and will be successively updated. The main purpose of DATED is to serve as a source for the deglaciation history of the Eurasian ice sheets to provide accurate digital maps with calendar-year isochrones to modelers and other researchers. It will also serve as an aid for identifying areas where data is lacking, and to facilitate future reinterpretation of the deglaciation pattern.

Key words (GeoRef Thesaurus, AGI): glacial geology, ice sheets, deglaciation, geographic information systems, data bases, dates, DATED Project, Eurasia.

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INTRODUCTION

The timing and pattern of the deglaciation of the Eurasian Ice Sheet is of key importance for Late Quaternary environments, on a local and global scale. The increasing resolution of glacier and climate models demands detailed information about the deglaciation on a calendar-year time scale. This is the rationale for the ongoing project DATED, whose goal is to serve as a primary source of information about the ice growth towards LGM and the deglaciation of Eurasia by presenting digital maps of ice growth and decay, and a database of published dates (^{14}C, OSL, TL, cosmogenic exposure, clay-varve), end moraines, and other relevant features.

Large volumes of data on the Eurasian deglaciation pattern and chronology exist in the literature, but most investigations provide information only on the local or regional level. Therefore, it is difficult to make wide ice-sheet syntheses, as interpretations between areas differ, and much of the relevant information is published in local journals, maps, and theses. Previous large scale reconstructions of the entire Eurasian Ice Sheet or its major constituents (e.g. Andersen 1981, Lundqvist & Saarnisto 1995, Kleman et al. 1997, Boulton et al. 2001, Svendsen et al. 2004) have the disadvantage of not being fully transparent to their data sources. These reconstructions are therefore difficult to validate and refine by further investigations. Recently, large efforts have been made to overcome these problems by creating digital maps connected to relational databases, where the information can be traced back to its original source. The GIS maps produced by Ehlers and Gibbard (2004) shows a compilation of glacial morphological features from the entire Eurasian Ice Sheet, but the level and amount of information varies widely between the contributing countries, and chronologic information is only available as sparse, interpreted ages of certain features, often without reference to the original dates. The BRITICE-project (Clark et al. 2004, Evans et al. 2005) produced a freely available GIS with detailed information about glacial features related to the last British (Devensian) Ice Sheet, with full source-transparency. The BRITICE database, however, does not include chronologic data, partly because of the very limited number of available dates (Clark et al. 2004). The success of the BRITICE project was a major inspiration for creating the ice-sheet wide DATED database, where all available morphologic as well as chronologic information is collected.

The main purpose of DATED is to serve as an updated source of interpreted ice margins and dates for the deglaciation of the Eurasian Ice Sheet, to provide accurate digital maps with calendar-year isochrones to modelers and other researchers. It will also show where data is lacking, and facilitate reinterpretation of the deglaciation pattern.

The present paper is a project description of work in progress. The first version of DATED will be published and made available on the internet in 2007, in a format readable for ordinary web-browsers. Information about DATED can be found on the web site http://www.gyllencreutz.se. This task is a part of the British-Dutch-Norwegian co-operative project Ocean Reconstruction and Modelling of the European deglaciation (ORMEN, http://www.bridge.bris.ac.uk/projects/ORMEN/Index.html) led by Sandy Harrison, University of Bristol, and is also part of the project Ice Age Development and Human Settlement in northern Eurasia (ICEHUS), led by John Inge Svendsen, University of Bergen.

STUDY AREA

The area comprised in DATED includes all areas covered by the northwest Eurasian (Scandinavian-, British-Irish-, and Barents-Kara-) ice sheets (Fig. 1), that is part of or the entire following countries: Ireland, Great Britain, Denmark, Norway, Sweden, Finland, Germany, Poland, Ukraine, Belarus, Lithuania, Latvia, Estonia, Russia, the Arctic areas between 0° E and 110° E, as well as the surrounding sea areas.

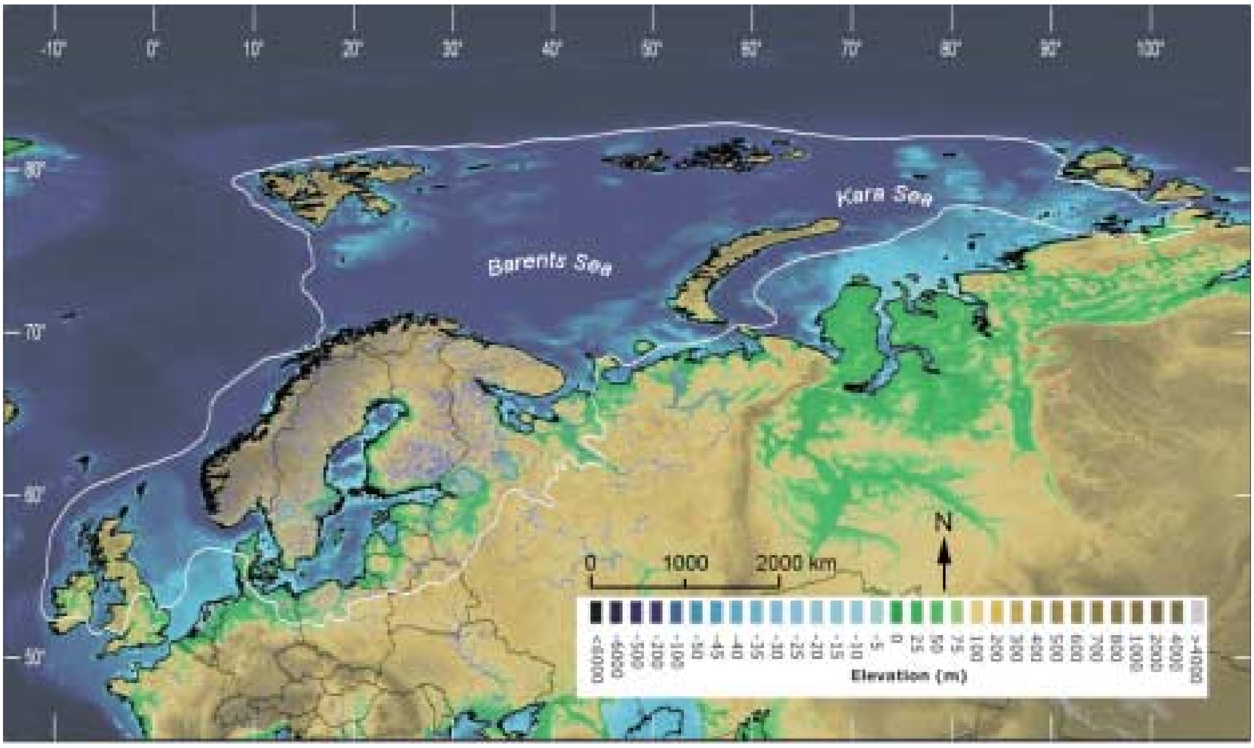

Fig. 1. Geographical extension of DATED, which is approximately within the asynchronous LGM reconstruction by Svendsen et al. (2004) (white line). Elevation data from the SRTM30PLUS dataset (Becker & Sandwell 2006).

METHODS

Database (DATED 1)

The database is created and maintained using Microsoft Access®. The goal of the database is to cover all available dates of the Eurasian deglaciation, based on radiocarbon (^{14}C), optically stimulated luminescence (OSL), thermoluminescence (TL), cosmogenic exposure (e.g. ^{10}Be, ^{36}Cl), and clay-varve records. The database will include ice growth towards the LGM, and therefore start with dates of the Ålesund Interstadial, c. 35-38 cal kyr BP (Mangerud et al. 2003). Dates showing the vertical extent of the ice sheet will also be included (e.g. Linge et al. 2006, Ballantyne et al. 1998).

The dates in the database are collected from published scientific papers of various kinds (original research articles, reviews and syntheses, map descriptions, or PhD-theses). When necessary, additional metadata (e.g. map coordinates and sediment descriptions) are acquired using the unique laboratory codes, for example, from published ^{14}C-laboratory reports (e.g. Håkansson 1982) or directly from the author.

All entries or changes to the database are performed by the authors of this paper to secure the data integrity and quality control. Researchers who wish to contribute with data or otherwise comment on the work are encouraged to contact us directly. The database will be maintained by the Department of Earth Science, University of Bergen until 2011, after which a permanent location for the database and maintaining personnel will be determined.

Data are entered into the database in two different ways. Single dates are entered through a custom made form, with fields and drop-down menus for all entries. Large numbers of dates from common source publications are entered together, to minimize the risk of entry errors, after customizing the table to match the DATED database structure. It should be noted that as a rule, only published information (including PhD theses) is used, which provides a crude quality control, and ensures that all data can be traced back to its original source. Exceptions to this are made for unpublished dates that supply verifiable information sufficient for a critical evaluation (name of researcher, lab code, map coordinates, dated material, sedimentary setting); these dates are still traceable to laboratory records, and are entered (as "unpublished") into the database with reference to the researcher who supplied them.

The database contains a large number of potentially relevant entry fields, of which the following features are considered crucial for incorporation to the database: Country, Locality, Longitude, Latitude, Dated material, Stratigraphic relative position (relation to ice margin position), Dating method, Laboratory code, Age, Age error, References, and Flag. "Flag" denotes comments important for interpretation of the dates, for example, if a date was interpreted to be erroneous by the original or subsequent authors. For dates obtained from compilations such as reviews or syntheses, references are given both to the original publication and to the compilation. Only short references (author(s), year) are given in the database, while a complete list of full references will be given in a separate document available from the same web site.

All dates in the database are also given in a format considered to represent calendar years. Radiocarbon dates are entered as uncalibrated ages, but are also given as (re-) calibrated dates using INTCAL04 (Reimer et al. 2004). Marine samples are corrected for reservoir ages using ΔR-values given in the database and Marine04 (Hughen et al. 2004) for the Holocene. For older marine dates, we will use reservoir corrections from Bondevik et al. (2006) and other new results. No calibration is performed on OSL, TL, or Exposure dates; these are treated as calendar ages. Clay-varve records connected to the Swedish Time Scale (STS) are corrected by adding 900 years to varve ages younger than 10,300 varve years BP, based on the correlation between the STS and the GRIP $\delta^{18}O$ record by Andrén et al. (1999, 2002) in conjunction with the new Greenland ice core chronology GICC05 (Rasmussen et al. 2006). The database is coupled to the GIS (DATED 2), enabling rapid data viewing by selecting features in the map view (Figs. 2 and 3), as well as advanced querying.

GIS (DATED 2)

The GIS maps are created using Intergraph's GeoMedia® Professional software. Besides basic cartographic information (country borders, lakes, rivers etc.), the DATED GIS contains a digital elevation model (DEM), mapped and interpreted ice margins, ice marginal deposits (end moraines, glacifluvial deltas, eskers, hummocky moraines, and boulder fields), and interpreted isochrones, with references to original authors as well as to authors of compilations, reviews or syntheses. Examples of the GIS and the database data are shown in Figs. 2 and 3. More detailed examples with references will be presented when we first publish the results of DATED. For practical reasons, no references for the data in Fig. 2 (except for the highlighted example features) are given here, because a full reference list would be too space consuming. Reconstructions of the vertical extent of the ice sheet will also be included (e.g. Linge et al. 2006, Ballantyne et al. 1998). DATED is also planned to include ice-dammed lakes and

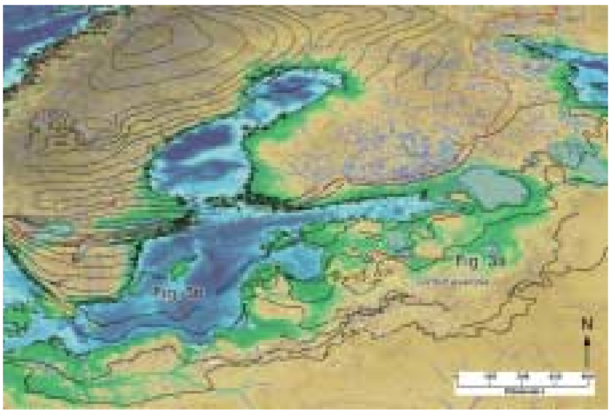

Fig. 2. Detail of the DATED GIS, showing the southeastern sector of the Scandinavian ice sheet and a sample of the data incorporated by October 2006. Black lines indicate ice margins, and white dots mark locations of dates. All features are thoroughly referenced in the databases, but for practical reasons, no references for the data sources are given here. The dark red line indicates the Younger Dryas ice margin according to Ehlers and Gibbard (2004). Examples of the GIS and database entries are given in Fig. 3. The yellow circle shows an example of (unedited) conflicting ice margin locations within the North Lithuanian Moraine (western line by Zelcs and Markots (2004) and eastern line by Raukas et al. (1995)).

ice-directional indicators (meltwater channels, eskers, drumlins, lineations and striae), inspired by the BRITICE project (Clark et al. 2004), but the chronology and pattern of the deglaciation is prioritized. The GIS map features are generalized for the scale 1:2.5 million.

Two different modes of workflow are used for the primary data capture. For the incorporation of data from previous GIS sources, such as the GIS CD-ROM in “Quaternary Glaciations - Extent and Chronology Part I: Europe” by Ehlers and Gibbard (2004), digital ice margins are imported to the DATED GIS in their original format, and additional metadata (e.g. references, ages and alternative ice margin names) are added. Published deglaciation maps or map figures are scanned and then georeferenced in the GIS using affine transformation (most often based on control points along the coastline), before the ice margins are digitized into the GIS, and additional metadata (as above) are added.

a. Database record example

Name	Value
ID	36
Country	Estonia
Region	
Locality	Kaagvere 2
Lon	26.6122222222222
Lat	58.005
Elevation_masl	
Depth_m	
DatedMaterial	Fine sand
StratigraphicUnit	
StratRelativePosition	Betw. 2 uppermost tills
DatingMethod	OSL
LabID	TlnOSL-1426
SampleName	
14CAge	
14CAgeError	
Age	14930
AgeError	1790
CalAge	
CalAgeError	
CalAge1sigmaLower	
CalAge1sigmaUpper	
Ref	Kalm (2006)
CompilationRef	
Flag	

b. Ice margin record example

Name	Value
ID1	1
IMZ_name	South Middle Bank Phase
LocationRef	Mojski (2000); Uscinowics (1999)
DatingMethod	14C
AgeRaw	12.7 - 12.6
AgeCal	14.9 - 14.6
AgeError	
AgeRef	Kalm (2006)
Country	Baltic Sea

Fig. 3. Examples of the information stored in DATED, as displayed in the GIS when selecting a) a date (white dots in Fig. 2), or b) an ice margin (black lines in Fig. 2).

CONFLICT MANAGEMENT OF ICE MARGIN LOCATIONS

Several areas show conflicting ice margin locations due to previous digitizing errors, gaps or misfits between maps by different authors, or due to inconsistent interpretations (Fig. 2). In these areas, the original lines are edited to construct a consistent and glaciologically plausible pattern, and the modified lines are stored separate from the originals in the GIS, with thorough referencing. The editing is performed in collaboration with the relevant authors, with the additional aid of stratigraphic information, satellite images, and high-resolution DEMs (c.100 m resolution).

DISCUSSION

Definition of the Last Glacial Maximum (LGM)
One goal of DATED is a map of the isochronous ice margin location at the time when the integrated ice covered area was the greatest, often called the LGM. However, the LGM is not an unambiguous term. The LGM ice margin, as reconstructed by Svendsen et al. (2004), Boulton et al. (2001), Kleman et al. (1997), and Andersen (1981) for example, represents the furthest ice margin advances between about 30 and 15 kyr BP (Houmark-Nielsen & Kjaer 2003), and not an isochronous ice margin. Since the timing of the maximum extent of the ice margin depends on glacier dynamics and regional variations in climate, topography, and bed conditions, the ice margin was at its maximum position at different times along the margins of the Eurasian Ice Sheet, and certainly between different glaciers and ice sheets. A definition of the LGM as the most recent maximum in global ice volume at about 21 kyr BP (Mix et al. 2001) is thus not applicable to the areal extent of individual glaciers or parts of ice sheets, and the former should, in our opinion, be called the Last Global Glacial Maximum (LGGM). A map of isochronous ice margins during the extended time period of ice advance and retreat, will demonstrate the asynchroneity of the "LGM-line" of Svendsen et al. (2004) for example, and provide a way to determine the maximum areal (and volume) extent of the Eurasian Ice Sheet. This can eventually be conceived through spatial querying of the DATED GIS isochrones, given sufficient data input.

CONCLUSIONS

1. DATED is an ongoing effort to produce an improved and transparent reconstruction of the last growth and deglaciation of the Eurasian ice sheets (Barents-Kara, Scandinavian, British). The first version will be published on the web in 2007 and present ice margins and dates around the entire perimeter of the ice sheets in a calendar-year scale. The data used is compiled from the literature, with references to all sources. DATED consists of two parts.
2. Part one is the Database on Eurasian Deglaciation Dates (DATED 1) was designed to cover all available dates (^{14}C, OSL, TL, cosmogenic exposure, and clay-varve records) of the last growth (from c. 40 cal kyr BP) and decay of the Eurasian ice sheets.
3. Part two is the Digital Atlas of the Eurasian Deglaciation (DATED 2) presenting the new reconstruction. However, DATED 2 also includes a database of end moraines, other morphological features and earlier reconstructions of the deglaciation for the entire or parts of the area. On the web, DATED 1 and 2 are connected so that readers easily can identify the background data for reconstructions.
4. The main objective with DATED is to serve as an updated source of interpreted ice margins and dates for the deglaciation of the Eurasian Ice Sheet, to provide accurate digital maps with calendar-year isochrones to modelers and other researchers. The transparency to data sources will facilitate evaluation and reinterpretation of the deglaciation reconstruction. DATED may also assist field workers in locating future areas of investigation, by showing where data is lacking or sparse.

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TIME-TRANSGRESSIVE EVOLUTION OF LANDSLIDES POSSIBLY INDUCED BY SEISMOTECTONIC EVENTS IN LAPLAND

by
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Sutinen, R., Piekkari, M. & Liwata, P. 2007. Time-transgressive evolution of landslides possibly induced by seismotectonic events in Lapland. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 121–128, 7 figures.

Paleolandslides are evidence of the occurrence of seismotectonic events associated with glacio-isostatic recovery, but their time-transgressive distribution is poorly understood in northeastern Fennoscandia. We investigated Early Holocene paleolandslides along a gradient from Utsjoki close to the Younger Dryas End Moraines (YDEMs) in northern Finnish Lapland down to Kittilä, close to the so-called ice divide zone of the Fennoscandian Ice Sheet (FIS) in southwestern Lapland. Morphological evidence demonstrated that the Utsjoki landslides were nunatak/ice-marginal meaning that during the slides the valley floors were still ice-covered, and we estimated that the slides had occurred around 10,180 yr BP. Paleolandslides in Kittilä are younger and yielded ^{14}C-ages up to 8,700 yr BP. According to present knowledge, at the northeastern Fennoscandian paleolandslides, major fault instability had been time-transgressively present at and behind the retreating margin of the FIS after the YDEMs.

Key words (GeoRef Thesaurus, AGI): landslides, neotectonics, faults, seismotectonics, Holocene, Lapland Province, Finland.

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INTRODUCTION

Northern Fennoscandia has experienced high-magnitude (up to Mag=8.2) postglacial fault deformations (PGFs) attributable to glacio-isostatic rebound (Arvidsson 1996). According to the subsidence-rebound model by Stewart et al. (2000), the occurrence of PGFs includes bending of the crust causing stress at the edge of the receding ice sheets, hence the ice-marginal terrains were subjected to seismic impacts (Arvidsson 1996, Wu et al. 1999). Conceptual models suggest paleolandslides were associated with these crustal events (Kujansuu 1972, Lagerbäck 1990, Sutinen 1992, 2005, Olesen et al. 2004), but their time-transgressivity and distribution particularly in the region close to YDEMs have remained obscure.

On the basis of radiocarbon dates (^{14}C) of landslide-buried organic sediments, we have recently found that the PGF-activity had continued through the early Holocene in west-central Lapland (Sutinen 2005). This finding demonstrates that glacioisostatic recovery involving major earthquakes seems to have continued at least 2000 years after the disappearance of the FIS in western Finnish Lapland. If the time-transgression concept holds for late-/postglacial crustal rebound, paleolandslides close to the YDEMs in Utsjoki should be older compared to those in the so-called ice-divide zone in Kittilä (see Fig. 1). Even though recent fault instability is minimal (Milne et al. 2001) compared to that of late Pleistocene-early Holocene recovery events (Arvidsson 1996), reconnaissance of PGFs and paleolandslides provides important information for example, for site evaluations within nuclear waste disposal studies (Lagerbäck & Witchard 1983, Ojala et al. 2004). Our attempt was to i) document large-scale fields of landslides by acquiring data from terrains close to the YDEMs down to the so-called ice-divide zone of southwestern Lapland and ii) to evaluate if the northeastern Fennoscandian paleolandslides can be time-transgressively attributed to glacioisostatic rebound.

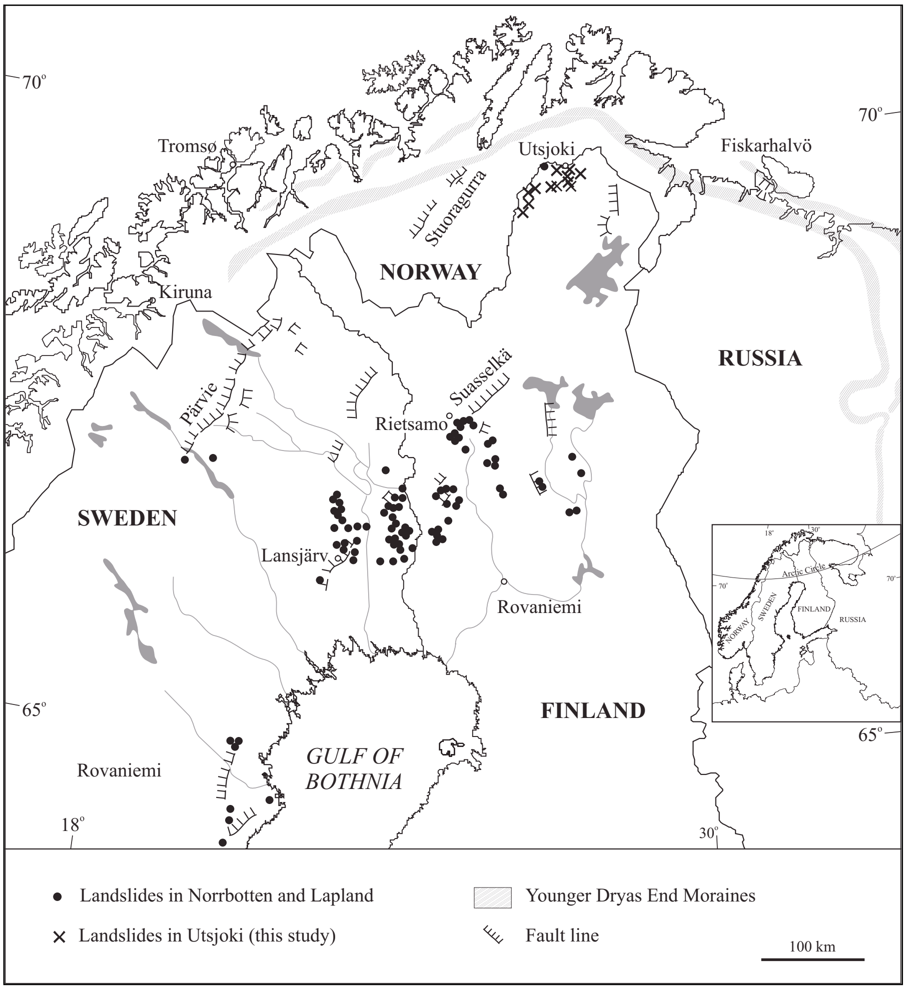

Fig. 1. Locations of northern Fennoscandian postglacial faults (PGFs) according to Tanner (1930), Kujansuu (1964), Lundqvist and Lagerbäck (1976), Lagerbäck (1979), Dehls et al. (2000), Sutinen (2005) and this study. Barbs along the fault lines are turned towards the lower block. Locations of late- or postglacial landslides according to Kujansuu (1972), Lagerbäck (1990), Sutinen (2005) (depicted as dots), and this study (depicted as crosses); Younger Dryas End Moraines (YDEMs; with grey shading) according to the Nordkalott Project (1987).

MATERIALS AND METHODS

We applied stereoscopic airphoto (AP) interpretation and digital elevation model (DEM based on the data-base by National Land Survey) to recognize landslides and neotectonic faults (tentatively PGFs) along a transect from Utsjoki 70°N28'E, 30 km south of YDEMs, down to Kittilä, 67°30'N24', Finnish Lapland (Fig. 1). Much of the conspicuous landslide scars in Utsjoki were found on the range of treeless fells (high altitude and strong relief features above treeline) that reaches 330-443 m (a.s.l.) in the Stuorra Skallovarri-Kanesvarri-area (Figs. 1–4). The complex of paleolandslides in Stuorra Skalluvarri covers an area of 30 km^3 (Fig. 2).

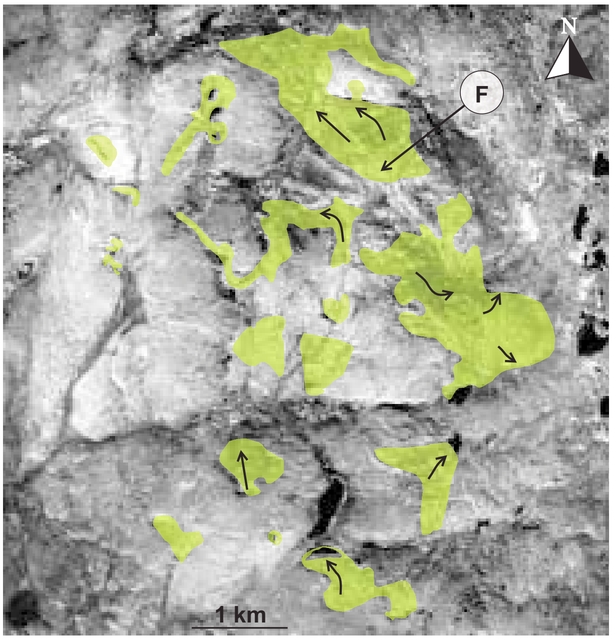

Fig. 2. Landslide scars (slide directions shown by arrows) and PGF-lines (shown as F) in the Stuorra Skallovarri fell area, Utsjoki.

The largest 'continuous' field of paleolandslide ridges and mounds found in Utsjoki is the Lake Fanasjavri slide covering an area of approximately 9 km^2, and located at an altitude of 400–360 m a.s.l. (Fig. 5). Field reconnaissance of the slides and faults were carried out in May 2005.

A number of rotational slip-type paleolandslides in Kittilä (Fig. 1) are well documented in earlier papers by Kujansuu (1972) and Sutinen (2005), hence we targeted our surveys on previously unrecognised field of slides, located in the Rietsamo area, western Kittilä (location in Fig. 1). This particular field of paleoslides covers more than 20 km^2 on the southern slope of the Rietsamo fell, reaching an elevation

Fig. 3. Down-slope view of a landslide scar (800 m long, 350 m wide) on the north-sloping (2%) Kanesvarri fell, Utsjoki. The slide thrust wall highlighted with mountain birch trees. Note the barren rock on the bottom of the scar.
Photo R. Sutinen 2005.

Fig. 4. Upslope view of landslide scar (750 m long, 400 m wide) on the south-sloping (8%) Kanesvarri fell, Utsjoki. Blocks up to 2 m^2 are common in the slide material.
Photo R. Sutinen 2005.

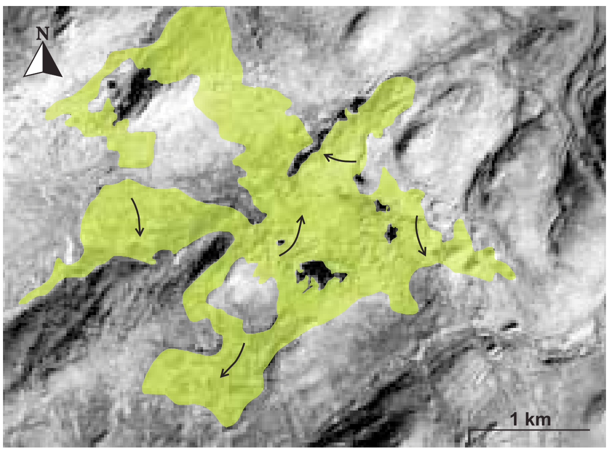

Fig. 5. Paleolandslides (slide directions shown by arrows) covering an area of 9 km^2 at Lake Fanasjavri site, Utsjoki.

of 310–250 m a.s.l. (Fig. 6). The highest peaks of fells in the area reach 425 m a.s.l. Within the routine mapping of Quaternary deposits (scale 1:400k) these landforms have been interpreted as hummocky moraines (without genetic specification) by Penttilä and Kujansuu (1963). Following the analogy of studies by Kujansuu (1972) and Sutinen (2005) that indicate that the fabric of the landslide materials are dissimilar to those of ice-flow directions, we attempted to rule out the concept of ice-deposition by determining the preferred orientation of the materials in cores of the landslide ridges in Rietsamo. Instead of conventional techniques (like excavation of test pits and fabric analyses), we determined the maximum anisotropy (orientation) of the sediments at 27 sites by applying non-destructive azimuthal measurements of apparent electrical conductivity (σ_a) (Taylor & Fleming 1988, Sauck & Zabik 1992, Penttinen et al. 1999). We applied a Geonics EM31 (Geonics Ltd. Mississauga, Ont. Canada) device designed to a depth range of 3 m in a horizontal coil position EM31(H) to determine the σ_a-anisotropy of sediments forming the cores of the landslide ridges. Reference data was acquired at four sites on the till plain south of the margin of the slide ridges. The field campaign was carried out in June 2006.

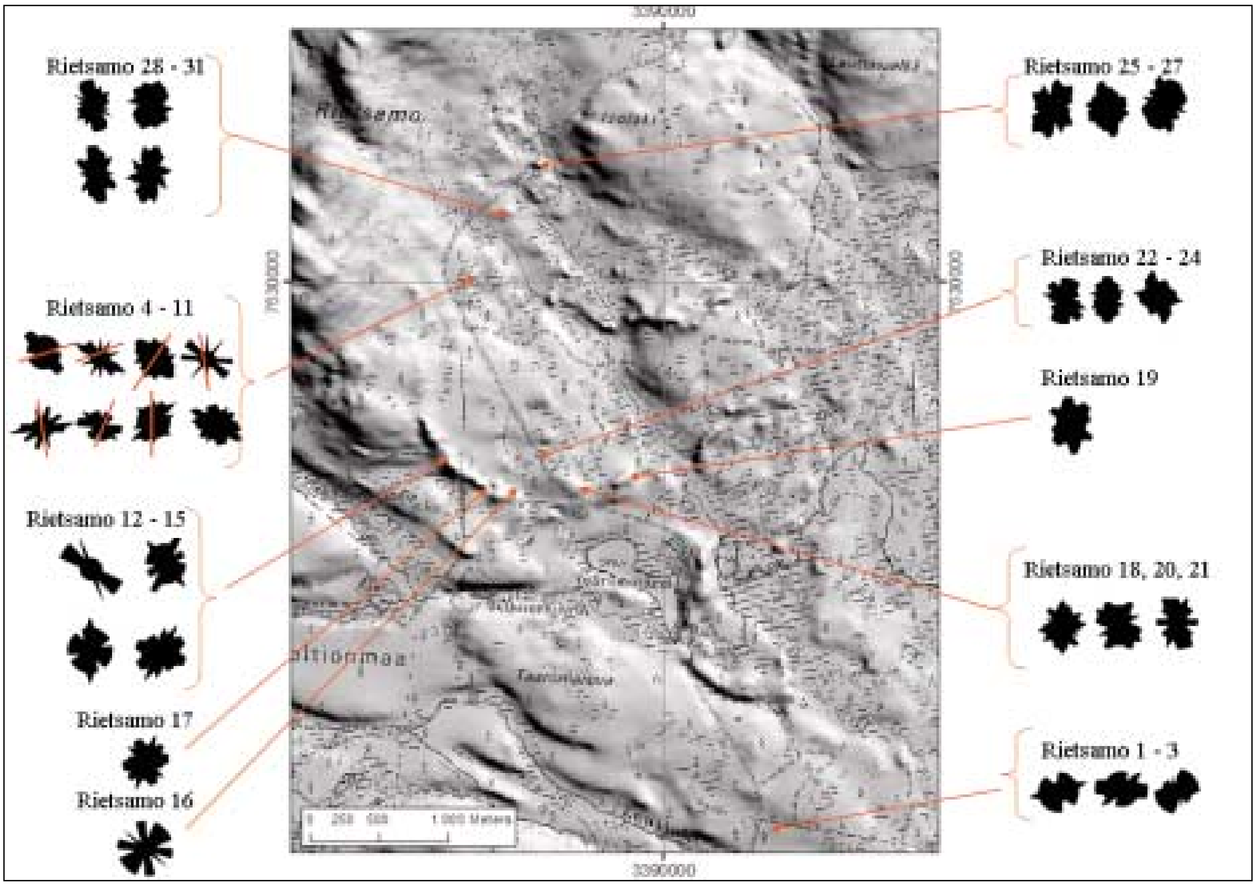

Fig. 6. DEM showing paleolandslide ridges fanning out to the south on the Rietsamo fell, western Kittilä. Sediment anisotropy shown as rose diagrams of azimuthal conductivity measurements with EM-31 in horizontal coil position. Elongation of small ridges (numbers 4-10) shown by red bars on rose diagrams. Field measurements conducted in June 2006.

RESULTS AND DISCUSSION

Postglacial fault alignment

The orientation of the AP-interpreted fault lines (tentatively PGFs) in Utsjoki (Figs. 1-2) was inconsistent with the predominant southwest-northeast pattern found in Kittilä and elsewhere in northern Fennoscandia, but is rather similar in orientation to those (NW-SE) reported from Kalastajasaarento (Rybachi Peninsula) (Tanner 1930) and those in the Kola Peninsula (Roberts et al. 1997). On the basis of a higher block in the down-ice direction, one may consider the northwest-southeast pattern of the PGFs to coincide with the retreat pattern of the margin of FIS after the YDEMs (see Norkalott Project 1987). The major PGFs recognized in northern Fennoscandia, however, are trending southwest to northeast; including the Suasselkä-Homevaara fault in Finland (Kujansuu 1964, Sutinen 2005), the Pärvie fault in Sweden (Lundqvist & Lagerbäck 1976, Lagerbäck 1979) as well as the Stuorragurra fault in Norway (Dehls et al. 2000; Fig. 1). Such orientation seems not to directly coincide with the retreat pattern of the FIS.

Morphology

We found more than 30 paleolandslides/landslide scars in the Utsjoki area (Figs. 1-5). The slides were found at altitudes of 405-360 m (a.s.l.), demonstrating that the valley floors were still ice-covered at the time of slides (Figs. 2 and 5). This is particularly well demonstrated in the case of the Lake Fanasjavri paleoslides at altitudes of 400-360 m a.s.l., but with no signs of sediment disturbances on the valley floor at an altitude of approximately 300 m a.s.l. (Fig. 5). The slide features on the Stuorra Gallovarri fell were distributed on all expositions of the fell slopes (Fig. 2), but some slides were found even in moderately flat surfaces. The length of the slides varied from 0.9 to 1.2 km, width from 0.3-0.5 km, and slope from 3.5 to 7.8%, respectively. The initiation of the slides had occurred at 390-405 m a.s.l. Bare rock and/or large boulders were often on the floors of the slide scars (Fig 3-4), demonstrating that soil moisture saturation had preceded the slide events (Sutinen 1992, 2005, Hänninen & Sutinen 1994).

Fig. 7. Outlook of small paleolandslide ridge (foreground/ clearcut; location shown by arrow for sites 4-11) and 15m high ridge (upper right-hand corner/ Norway spruce-downy birch-aspen-forest) in Rietsamo, Kittilä.

The Rietsamo field of paleolandslides (see location in Fig. 1) are composed of arch-shape ridges varying in height from 2 to 15 m and fanning out to the south (Figs. 1 and 6). Some of the ridges in the northern part of the landslide field (sites coded R25-27 in Fig. 6) were mantled with boulders, but the material, in general, was not coarse-textured. According to field observations and the composition of forest canopy, which included Norway spruce as a dominant tree species and downy birch as a principle associate (Fig. 7) with a silty matrix that was rather common in the slide materials (see Sutinen et al. 2002). However, texture or matrix does not play a crucial role in sliding susceptibility, but a much more important feature is soil saturation as seen with modern and Holocene landslides (Keefer 2002, Sutinen 1992, 2005, Hänninen & Sutinen 1994, Friele & Claque 2002).

Sedimentary anisotropy

Azimuthal measurements of electrical conductivity (σ_a) with EM31(H) indicated that the slide materials in the Rietsamo ridges had a maximum anisotropy indicating a fanning-out pattern to the south (Fig 6; R16-31). This pattern clearly deviated from the maximum anisotropy (WSW-ENE) obtained from the reference till plain on the south side the slide field (Fig. 6; R1-3) or R12 that was made (distally) just outside the landslide ridge demarcating the Rietsamo field of paleolandslides (see Fig. 6). The crests of small ridges such as those shown in Fig. 7 (out of DEM-resolution) were oriented as follows: 60-260° for R4-5, 30-210° for R6, 170-350° for R7-8 as well as 20-200° for R9, 0-180° for R10. R11 was on a non-oriented hummock. The variation in maximum σ_a-anisotropy was large, suggesting that the sliding event(s) was catastrophic in nature and likely triggered by fault instability. The anisotropy-pattern was not coherent with the ice-flow stages, which on the basis of till fabric have the southwest-northwest (Late Weichselian) and northwest-southeast directions (Early Weichselian) (Sutinen 1992). The nearest known neotectonic fault line is the Homevaara fault, 13 km southeast of Rietsamo landslide field (Sutinen 2005). Homevaara fault is an extension of the Suasselkä fault described by Kujansuu (1964, 1972).

Time-transgression

The age of the YDEMs, close to the border of Finland and Norway (Fig. 1), has been estimated to be within the envelope of 12,500-10,300 yr BP (Sollid et al. 1973, Andersen et al. 1995). Applying the YD (ended 10,300 yr BP) ice surface model by Andersen et al. (1995) and the descend rate of 1.25 m/yr of the ice surface, as we estimated with ice-marginal (annual) meltwater channels, the studied landslides occurred around 10,180 yr BP in Utsjoki. The retreat rate of the post-YD ice margin was estimated about 80-90 m/yr, hence the area was ice-free by around 9,900 yr BP. The above time frame estimated for the Utsjoki paleo-landslides is concurrent with the maximum fault instability in Fennoscandia between 11,000-9,000 yr BP (Wu et al. 1999).

The time-transgressivity of the landslides inside the YDEMs is attributed to maximum crustal subsidence in the interior part of Fennoscandia (Fjeldskaar et al. 2000; Stewart et al. 2000) and subsequent fault instability within the glacio-isostatic recovery (Arvidsson 1996, Wu et al. 1999). Consequently, the mass-flows of the offshore sediments in the Norwegian fjords provides evidence of at least three separate large-magnitude earthquakes affecting northern and western Norway during the period of 11,700-9,000 yrs. BP (Bøe et al. 2004, Olesen et al. 2004).

No morphological evidence was found to indicate that the Rietsamo paleolandslides in Kittilä were ice-marginal. Hence, those likely fall into the time-envelope of 9,300-9,000 yr BP based on stratigraphic evidence for the intra-plate PGFs in the Nordkalott area (Lagerbäck 1992, Dehls et al. 2000), and/or 8,140-6,600 yr BP based on indirect age-estimates from peat deposits on top of the slide scars (Kujansuu 1972, Lagerbäck 1990). The best estimate so far are the ^{14}C-dated fragments of subfossil trees found beneath the landslide materials yielding age of 8,720±170 yrs B.P. (cal 9,730 yr BP) in Kittilä (Sutinen 2005). These observations indicate the paleolandslides in Kittilä are younger than those in Utsjoki and the fault instability associated with glacio-isostatic recovery was evidently present at and behind the retreating ice margin of the FIS after the YDEM in a time-transgressive manner.

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Geological Survey of Finland, Special Paper 46, 129–134, 2007.

GEOLOGICAL EVIDENCE OF THE SEISMICITY IN THE KOLA REGION DURING THE LATE PLEISTOCENE AND HOLOCENE

by
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Yevzerov, V. & Nikolaeva, S. 2007. Geological evidence of the seismicity in the Kola region during the Late Pleistocene and Holocene. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 129–134, 2 figures.

Deglaciation of the Kola region during the Late Valdaian (Weichselian) occurred mainly by separation of the vast ice peripheral covers from the active ice massif and the subsequent melt of dead ice. After last ice advance in the Younger Dryas, the ice sheet was dissected by extended marine gulfs during the beginning of the Holocene and quickly melted. The ice had the highest surface gradient in the Older Dryas and deglaciation was fastest in the area that was covered by the Younger Dryas ice. Glacioisostatic uplift of the Kola region in the western part of the Kola Peninsula was more intensive compared to the eastern part. Such a tendency prevailed during the end of the Late Pleistocene and Holocene. At the same time, the speed of glacioisostatic uplift decreased everywhere in time, but the uplift proceeds even at present.

Palaeoseismic dislocations studied using geologic-geomorphologic methods are expressed in the topography in the form of seismic ditches, scarps, gorges, fractures in crystalline rocks and disturbances in the Quaternary sediment units. Epicentres of the ancient earthquakes basically coincide with epicentres of modern earthquakes and are concentrated on territory that was covered by active ice with the greatest gradient of its surface during the Older Dryas and, which elevated most intensively after deglaciation of the area. All the data assumes that the main reason for the seismicity of the region is a tension that was caused the glacioisostatic uplift during and after deglaciation.

Key words (GeoRef Thesaurus, AGI): glacial geology, deglaciation, neotectonics, glacial rebound, seismicity, Holocene, Pleistocene, Kola Peninsula, Russian Federation.

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INTRODUCTION

Subsidence of the Earth's crust took place during glaciations, resulting from an increase in ice loading. The subsidence was followed by uplift during and after deglaciation. The glacial loading and unloading results a deep-seated viscoelastic response of the crust and produces faulting and earthquakes in its upper part (Stewart et al. 2000). The present article focuses on the seismic activity of the Kola region, which, in our opinion, is connected to deglaciation of the Late Valdaian ice sheets. The Murmansk district and adjacent areas of the shelves of the Barents and White seas are also discussed.

DEGLACIATION AND GLACIOISOSTATIC UPLIFT

In the Kola region, the Late Valdaian ice sheet reached its maximum extent 16,000–17,000 years BP. It had a low gradient surface during the maximum extent and the whole period of deglaciation. A low surface gradient was caused by an occurrence of peripheral parts of the ice cover on a low plain covered by friable water-saturated sediments similar to those lying on the modern continent and shelf. The presence of the water-saturated friable substratum on the glacier bed essentially reduced the resistance of ice movement at the contact between ice and bed and resulted in a slight inclination of the glacier's surface. Degradation of the ice sheets took place during cyclic climatic variations of different ranks (cf. Alm & Vorren 1990, Lehman & Keigwin 1992). Each variation took place within 500 to 2000 years including a rather fast warming and a subsequent gradual cooling. During interstadial warm events, the surface gradient of glacial ice was low and extensive peripheral ice margins were cut off from the main active ice mass (Yevzerov 1998). During the same warm periods, the edge of the dynamically active ice was in contact with rising or already existing periglacial water basins under favourable geomorphological conditions. The high heat capacity of water promoted a fast juxtaposition of the active ice margin with the line of zero glacier mass balance and a long preservation of the stable ice margin's position. Conditions for accumulation of thick glaciofluvial deposits were created at the ice front that terminated in water. Subsequently, after a complete disappearance of ice, these deposits are shown in the relief as dump moraine ridges, the so-called marginal eskers. dead ice melting was ongoing for a long time including several interstadial warmings. The glacier that advanced again during the phases of stadial coolings deformed interstadial deposits and built mainly push moraine ridges. Thus, a marginal belt was formed during each interstadial-stadial climatic cycle at the edge of active ice, which consists of two bands of marginal ridges: the inner band – a dump moraine, and the outer band – a push moraine. Three marginal belts (I, II, III) were formed during the period between the Last Glacial Maximum and the Preboreal in the Kola region. Based on the results of palaeogeographic research, the formation of belt III occurred in connection with the warming that took place between 14,700 and 16,100 years BP and the subsequent cooling in an interval from ~14,700 up to 13,400–12,900 years BP. Formation of marginal belts II and I correspond to climatic rhythms of the Bølling - Older Dryas and Allerød - Younger Dryas, respectively (Yevzerov 1998). The ice surface in the Older Dryas had the greatest ice surface gradient. After the last ice advance in the Younger Dryas, the ice sheet was dissected by extended marine gulfs during the warming at the beginning of the Holocene. In these gulfs, glaciomarine sediments were followed by marine deposits, probably due to the final ice melting about 9,000 years BP. The deglaciation was fastest in the territory occupied by the Younger Dryas ice cover.

Glacioisostatic uplift of the Kola region was a consequence of deglaciation. Isolation basins were studied in locations limited in area at different heights between the high marine limit and the modern seacoast. The deposits of marine- lacustrine or lacustrine-marine transitions have been studied in holes' cores, with subsequent diatomic analysis. Samples from transitional zones were dated using the radiocarbon method. Diagrams of the sea coastline movement were then reconstructed from the dating results and high-altitude positions of isolation basins thresholds. The work on the three locations on the Barents Sea and the three locations on the White Sea coasts of the Kola Peninsula were completed. The diagrams allowed for the construction of isobases for the territorial rise. The arrangement of isobases for the last 9,000 years and all the locations studied are shown in Fig. 1.

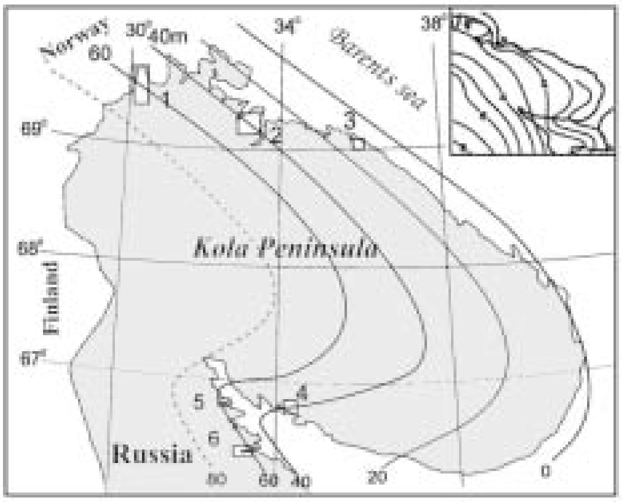

Fig. 1. Scheme isobases of the Kola region for the last approximately 9,000 years (on the data V. Yevzerov, V. Kolka, J. Møller, D. Corner; Snyder et al. 1997) 1-6 sites of investigations. On setting a modern rising of the Kola region in mm\year is shown (Andersen & Borns 1994).

This rise most likely had a dome-shaped form; the western part of the Kola Peninsula rose more intensively than the eastern one. Such a tendency prevailed during the end of the Late Pleistocene and Holocene. At the same time, the speed of uplift decreased in time everywhere, but nevertheless the raising continues today as well. It is not difficult to see that the nature of the rise has not changed until the present, which probably indicates that glacioisostatics has been present up to now. Geological data and theoretical calculations also testify to this suggestion. There is reason to believe that the territory examined should still be lifted by approximately 40-60 m. Buried mouths of big river valleys are located at the negative marks of such an order. These valleys were generated probably during the Pliocene. Their present position is most likely caused by the fact that the isostatic balance had no time to be restored during the rather short intervals of the Quaternary interglacials as the constant relaxation considerably exceeded the generally accepted one for several thousands years.

SEISMICITY

The Kola Peninsula, as an element of the northwestern Baltic Shield, is not a high-seismicity region at present. This has not always been the case as is seen in palaeoseismic deformation structures reflecting strong earthquakes. These were observed during the geological-geomorphological investigations and the analysis of air photos (Nikolaeva 2001). Palaeoseismic deformations are expressed in the topography as seismic ditches, scarps, gorges, fractures in crystalline rocks and disturbances of Quaternary sediment structures. These dislocations are accompanied by such phenomena as collapses, rock pillars, rock falls, empty niches (blocks displaced away) and other seismic signatures, which are not found outside seismogenous structures, and are confined to active fault zones. Deformations in incoherent sediments provoked by modern earthquakes are known not only in seismoactive regions, but also in territories of moderate seismicity, such as Canada, Sweden, Norway and Finland.

Recently, special studies have been carried out to define different types of deformations in Quaternary sediments exposed in quarries and natural outcrops along river banks in the northwestern Kola region and to discriminate these sediments from those related to glaciotectonic structures. All palaeoseismic deformations were formed during the late-glacial and post-glacial periods. Previously, we established that a strong earthquake occurred in the vicinity of Murmansk about 8,950±150 ca BP (Yevzerov & Nikolaeva 1995). This age value characterizes the initial stage of organic matter accumulation in a lake formation that was related to the damming of a small creek valley by landslides after the earthquake. A similar age for a seismic event (8,720±170 ca BP) was obtained in Finnish Lapland (Sutinen 2005). Seismogenic deformations were found in postglacial marine sediments in the Pechenga river valley (Nikolaeva 2006). The age of this seismic event was determined using the radiocarbon method on sea shells. The deformations occurred earlier than 8,500-8,700 ca BP. This period corresponds to the Boreal warm climatic phase that excludes any influence of glaciers or cryogenic processes on the deformation of sediments. Thus, the available data indicates that at least two seismic events occurred in the Holocene. Characteristic features of the morphology, dimensions and types of the palaeoseismic deformations allow us to conclude that the intensity of earthquakes (I) was no less then 7 and possibly 8 or greater according to the MSK-64 scale.

A scheme of ancient earthquake epicenters (Fig. 2) was made on the basis of geological data and air photo interpretation. Seismic dislocations directly indicate residual tectonic deformations in the epicenter zones of ancient earthquakes, which are largely concentrated in the west and centre of the Kola Peninsula. The epicenters of ancient earthquakes and the epicenters of most historical and recent earthquakes coincide, which testifies to the inherited nature of historical and recent earthquakes. The epicenters are concentrated in the area that was occupied by active ice during the Older and Younger Drays.

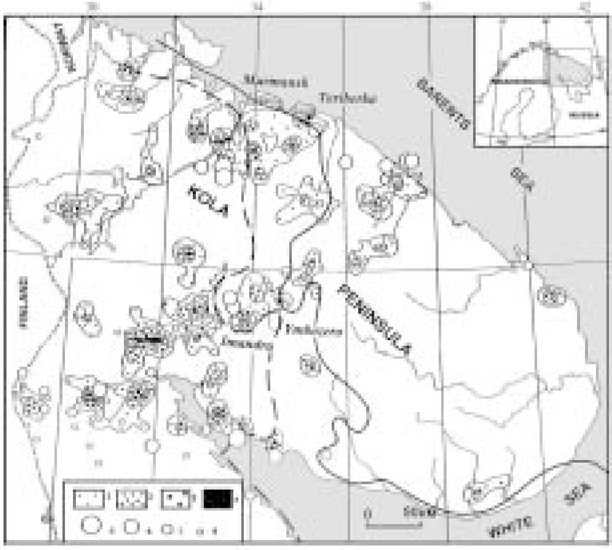

Fig.2. Scheme showing the density of residual deformations, the epicenters of modern earthquakes and the disposition of the edge of active ice in the Kola region. Residual deformation density distribution scale (numerals indicate the number of residual deformations over an area of 15x15 km): 1 – 1-3, 2 – 3-5, 3 – 5-7, 4 >7; epicenters of 1542-2002 earthquakes, magnitude M: 5 - ≥5.1; 6 -4.1-5.0; 7 -3.1-4.0; 8 - ≤3. The disposition of the edge of active ice is shown by a dotted line (during the Younger Dryas) and continuous line (during the Older Dryas).

The existing knowledge of seismicity of the region corresponds to a short-term period of instrumental and macroseismic observations. The instrumental measurements have been carried since 1956 when the seismic station “Apatity” started to work. For this period, the region was subjected to earthquakes with intensities of 3-4 and locally up to 5. Seismic activity zones are situated in the Kandalaksha Gulf, on the Murmansk Coast, especially in the Kola Bay area, and in the Khibiny Mountains. Information about historical earthquakes is recorded in archival documents, annals and newspaper publications. Such events have been known on the Kola Peninsula since the 17th century. The strongest historical earthquake took place in the Kandalaksha Gulf in 1627 with intensity 8 on the MSK-64 scale. It is possible to assume that the seismicity of the study region was, at that time, substantially higher than today.

CONCLUSION

The profile of the Late Weichselian ice sheet had a low gradient in the Kola region. Therefore, an areal type of deglaciation dominated for a long time. A dissecting type of deglaciation took place only in the Holocene. The epicenters of ancient and modern earthquakes mainly occupy the territory that was covered by active ice during the last stages of deglaciation and was glacioisostatically uplifted with the greatest speed after deglaciation. Such a distribution of dislocations is probably explained by the highest ice surface gradient of the area that was covered by active ice during the Older Dryas and by fast deglaciation of the territory occupied by active ice during the Younger Dryas. The general character of the glacioisostatic uplift has not changed until now and the elevation continues though its speed was considerably reduced. Modern and historical earthquakes are mainly inherited. All these data suggest that the main reason for the seismicity of the region is stress resulting from the glacioisostatic uplift of the Earth’s crust during the deglaciation and postglacial period.

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LITHOLOGICAL CONTRIBUTION TO HYDRAULIC PROPERTIES OF TILLS: INTER-SEASONAL DIELECTRIC ASSESSMENT

by
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Sutinen, R., Piekkari, M., Tepsa, A., Virtanen, M. & Hänninen. P. 2007. Lithological contribution to hydraulic properties of tills: inter-seasonal dielectric assessment. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 135–140, 5 figures.

Composition, particle-size distribution and clay mineralogy of tills are associated with the lithology of source rock types and the mode of deposition beneath continental ice sheets. The influence of lithological characteristics on hydraulic properties, particularly on the spatio-temporal variation of water content in tills, however, has remained obscure. Inter-seasonal monitoring of the dielectric properties of tills (ε_{TILL}) with varying composition in lithology and particle-size distribution was carried out from 1995 to 2003 in Finnish Lapland. Two sites were selected on the basis of similarity in climatic conditions, but dissimilarity in lithology (felsic granite gneiss-mafic chlorite-amphibole schist) and texture (sandy till-silty till), respectively. Inter-annual monitoring demonstrated that the variation in unfrozen soil water content of tills followed climatic events, but snowmelt was the major contributor to the intra-seasonal ε_{TILL}. The increase in ε_{TILL} was initiated by meltwater released from snowpack from two weeks to more than three weeks before the disappearance of snow in both cases. However, the response of snowmelt to ε_{TILL} was clearly site and texture-specific, such that maximum ε_{TILL}-peak for sandy till occurred several weeks before ε_{TILL}-peak of silty till. The timing difference of a month was observed between the latest spring (1996) and earliest springs (2002, 2003). The intra- and inter-seasonal soil ε_{TILL} was significantly lower ($F=810$, $p=0$) in sandy till than that in silty till. Even the maximum ε_{TILL} of the sandy site was minor to minimum for the silty site. It was found that the mean inter-seasonal $\varepsilon_{TILL}=7.0\pm1.36$ applies to sandy till and $\varepsilon_{TILL}=24.4\pm6.56$ to silty till, respectively. We estimated (99% confidence) that the inter-seasonal mean of sandy till varies between $6.7<\varepsilon_{TILL}<7.4$ and that of silty till between $22<\varepsilon_{TILL}<26$, respectively.

Key words (GeoRef Thesaurus, AGI): soils, till, water content, dielectric properties, textures, meltwater, Lapland Province, Finland.

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INTRODUCTION

The vadose zone (field saturated) hydraulic conductivity (K_f) of tills is attributed to lithological features, such that high K_f is typical in tills with a sandy matrix in regions of felsic granitoids or granite gneisses, while low K_f is found typically in tills derived from mafic rocks of the greenstone belt in Finnish Lapland (Sutinen et al. 1998, Penttinen 2000). A significant correlation has been found between the dielectric properties and soil volumetric water content (θ_v) of glacial drift materials (Sutinen 1992). Also, a significant correlation has been found between dielectric values and log-K_f for tills (Sutinen et al. 1998, Penttinen 2000). Thus, dielectric measurements and monitoring reveal differences in water regimes and hydraulic properties of tills derived from various lithologies.

The soil θ_v is influenced by soil texture, structure and topographic curvature (Vachaud et al. 1985, Sutinen 1992, Hänninen 1997), but climatic factors primarily contribute to the temporal variation of intra- and inter-seasonal θ_v (Solantie 1987). Similar to agricultural fields, studies on forest sites have revealed that, although soil θ_v varies in time and with location, the pattern of spatial variability does not change with time when the observations are ranked according to the magnitude of the soil θ_v (Sutinen et al. 2007). This phenomenon is called as time-stability (Vachaud et al. 1985). In addition, the four-step linear dielectric (ε) mixing model, as presented by Hänninen (1997), implies that geotechnical resistance, like bearing capacity, will significantly decrease at full saturation of sediments, for example ε=30 ($\theta_v \approx 40\%$), but beyond this critical point bearing capacity will be lost, such as on the log-hauling roads (Hänninen & Sutinen 1994). Hence, the knowledge of the spatial and temporal variability of dielectric properties of tills provides an effective alternative to estimate hydraulic features of tills that can be used in geotechnical planning and in the timing of harvesting and transportation logistics within forestry.

MATERIAL AND METHODS

Previous studies have indicated a significant correlation between dielectric properties and soil volumetric water content (θ_v) of glacial drift materials (r_s=0.98, P<0.01, n=425), such that $\theta_v = -0.06+0.024\varepsilon-0.00037\varepsilon^2+0.000003\varepsilon^3$ (Sutinen 1992; Fig. 1), and a significant correlation (R^2=0.79) between dielectric values and log-K_f of tills (Sutinen et al. 1998, Penttinen 2000). Hence, the in situ measurements and monitoring of ε_{TILL} reveal differences in water regimes and hydraulic properties of till derived from various lithologies. We investigated tills with varying composition in lithology and

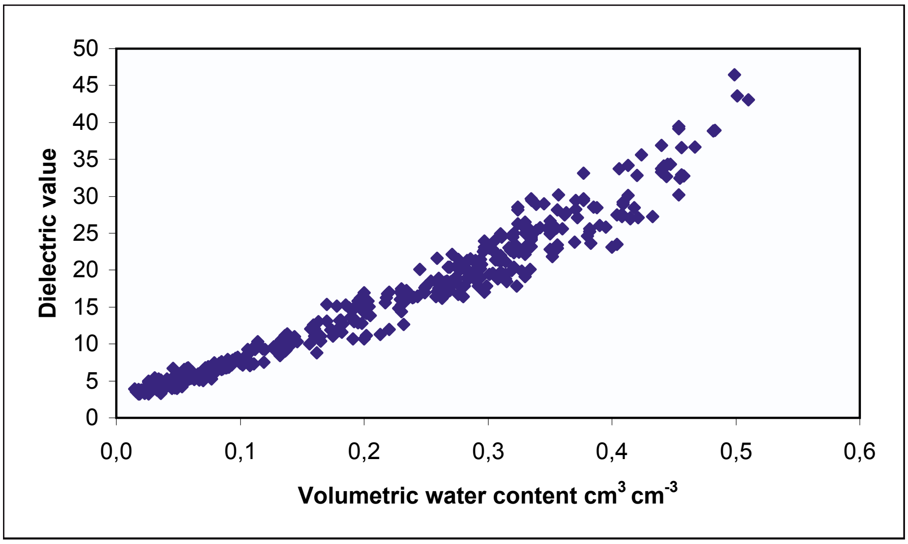
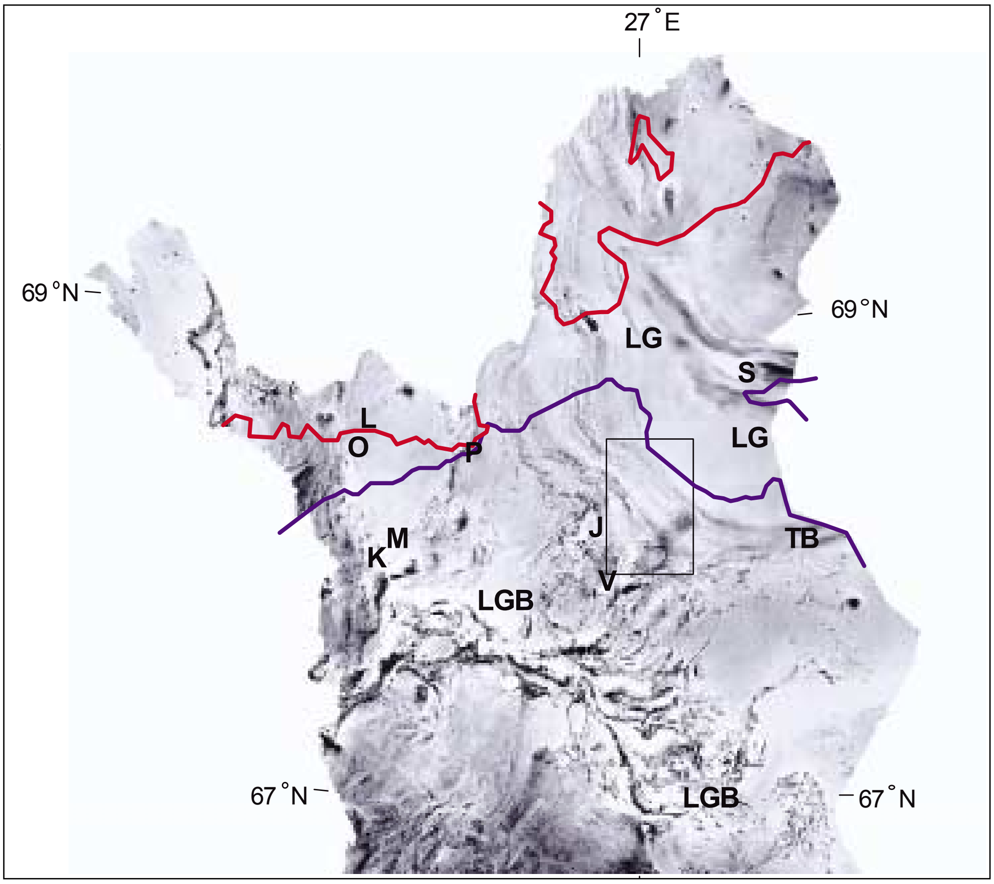


Fig. 1. A plot showing the correlation (rs=0.98, P<0.01, n=425) between soil dielectric values and volumetric water content according to Sutinen (1992). An empirical relationship $\theta_v = -0.06+0.024\varepsilon-0.00037\varepsilon^2+0.000003\varepsilon^3$ is obtained.

Fig. 2. Study sites Juolkusselkä (J) and Vaalolehto (V) in Finnish Lapland. LGB=Lapland greenstone belt, LG= Laplan granulite belt, TB= Tanaelv belt.



particle-size distribution in Finnish Lapland with inter-seasonal monitoring of dielectric properties. Two sites, Juolkusselkä (68°02'N and 26°26'E) and Vaalolehto (67°50'N and 26°41'E; see locations in Fig. 2), were selected for comparison on the basis of similarity in climatic conditions, but dissimilarity in lithology (felsic granite gneiss-mafic chlorite-amphibole schist; see lithology in Lehtonen et al. (1998)) and texture (sandy till-silty till; Fig. 3), as well as forest composition (pine-spruce), respectively. According to Haavisto (1983), fine-grained tills (silty tills) contain more than 5% of clay fraction and more than 30% of fines (i.e. silt and clay fraction).

To minimize sediment deformations during the monitoring, a permanent instrumentation of till sequences was applied. The design included transmission lines (parallel 15-cm-long steel probes with 6-cm-spacings) inserted horizontally into the till sequences (Sutinen et al. 1997). Recording of the dielectric values conducted with time domain reflectometer (TDR, Tektronix 1502B, Beaverton, OR, USA) on a weekly basis through 1995–2003.

Fig. 3. Silty till (fine-fraction content 52%) at the Vaalolehto study site.

RESULTS AND DISCUSSION

Texture effect

Inter-annual (period 1996–2003 shown in Figs. 4-5) monitoring demonstrated that variation in unfrozen soil water content of tills followed climatic events, but snowmelt was the major contributor to the intra-seasonal ε_{TIL}. The increase in ε_{TIL} was initiated by meltwater released from the snowpack over two weeks to more than three weeks before the snow disappearance in both cases. However, the response of snowmelt to the ε_{TIL} was clearly site and texture-specific, such that the intra- and inter-seasonal soil ε_{TILL} was significantly lower (F=810, p=0) in sandy till than in silty till. Even the maximum soil water content at Juolkusselkä site was minor to that of the minimum at the Vaalolehto site. It was found that mean inter-seasonal ε_{TIL}=7.0±1.36 applies to sandy till and v=24.4±6.56 to silty till, respectively. We estimated (99% confidence) that the inter-seasonal mean of sandy till varies between 6.7< ε_{TIL}<7.4 and of silty till between 22< ε_{TIL}<26, respectively.

The soil θ_v has been observed to be positively correlated with the fine fraction content in drift under forest stands (Sutinen 1992, Hänninen 1997, Salmela et al. 2001). Clay fraction content in tills in Lapland is rather low (<10%), but the fine fraction content of parent tills may reach 50% in LGB and TG, but may be less than 30% in LG (Hänninen 1997, Salmela et al. 2001; see lithological provinces in Fig. 2). The vadose zone (field saturated) hydraulic conductivity (K_f) of tills is attributed to these textural features. We have found that Kf of tills with sandy matrix falls into the range of K_f =10^{-5}–10^{-6} msec^{-1}, and is typical to regions of felsic granitoids or granite gneisses, while the range K_f =10^{-7}–10^{-10} msec^{-1} is found typically in tills derived from the mafic rocks of the greenstone belt in Finnish Lapland (Sutinen et al. 1998, Penttinen 2000). The consistent behaviour of ε_{TIL} indicates that the concept of dielectric time-stability is valid for the studied tills here (Sutinen et al. 2007, see Vachaud et al. 1985 for agricultural soils). This implies that fine-grained tills consistently and annually face spring saturation and loss of bearing capacity at sites with snow cover in winter. Hence, the reconnaissance of fine-grained tills is of major importance in the mapping of Quaternary deposits. Instead of conventional techniques (sampling and analysis of particle size distribution), the application of the low altitude airborne radiometric measurements provides an effective alternative for the mapping of tills with different water regimes (Hyvönen et al. 2003).

Snowmelt

The present results demonstrate that the rapid increase in the ε_{TIL} (soil θ_v) both in silty till at the Vaalolehto site and sandy till at the Juolkusselkä site were generated by melting snow. Our observations coincide with other studies from continuous permafrost regions of Alaska and Siberia (Boike et al. 1988, Hinkel et al. 1997, Kane et al. 2001) as well as from the Canadian subarctic (Leenders & Woo 2002) and northern Fennoscandia (Hänninen 1997, Sutinen et al. 1997). The inter-annual monitoring indicated that unfrozen soil water became available as early as three weeks before the snow disappearance in the tills of forest stands composed either of mature Scots pine or Norway spruce (Figs. 4-5). However, the till types responded to climatic conditions in different ways, and the onset timing of meltwater percolation varied between the years. Sandy till exhibited the maximum dielectric peaks systematically 3-4 weeks earlier as compared to those of silty till (Figs. 4-5). The time-difference between the latest (1996) and the earliest (2002 and 2003) spring saturation was one month for both types of tills.

Meltwater released from the snowpack can rapidly percolate down through soil pores or thermal contraction cracks (Hinkel et al. 1997, Kane et al. 2001) and infiltrate unimpeded into organic soil horizons (Leenders & Woo 2002). Thus, the fraction of unfrozen soil θ_v tends to increase markedly after the onset of the snowmelt (Boike et al. 1988). Similar to previous reports (Hänninen 1997, Sutinen et al. 1997), this study indicated that spring saturation, for example soil ε>30, θ_v>0.44 by cm^{3}cm^{-3} according to the dielectric mixing model as presented by Hänninen (1997), is an inter-annual phenomena for fine-grained drift materials derived for example, from chlorite-amphibolite schists and occupied by stands dominated by Norway spruce (Fig. 3). In contrast, snowmelt saturation is absent in sandy drift materials derived from granite gneisses for example, and covered by stands dominated by Scots pine (Hänninen 1997, Sutinen et al. 1997). In accordance with previous reports (Ruther 1999, Hagberg 2001, Salmela et al. 2001, Sutinen et al. 2002, Hyvönen et al. 2003), inter-seasonal monitoring of this study indicated that Scots pine was adapted to much lower soil water regimes than Norway spruce in Finnish Lapland. Pine is one of the best indicators of high permeability of tills.

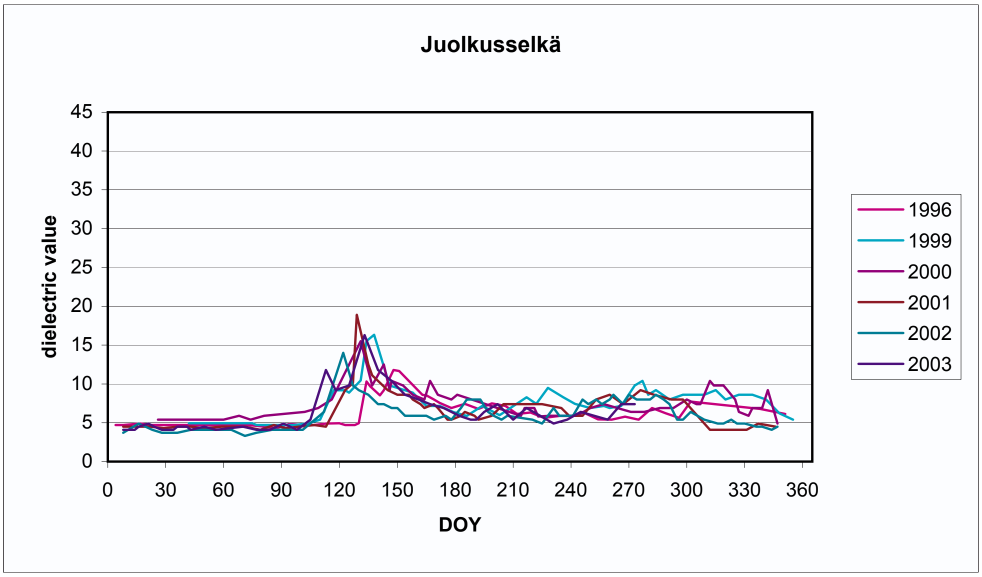


Fig. 4. Annual variation of the dielectric values of sandy till at the Juolkusselkä site presented separately for the years 1995-2003. Note difference between the onsets of water percolation through frozen soil (soil temperature <0ºC) in spring 1996 (May) and 2003 (April).

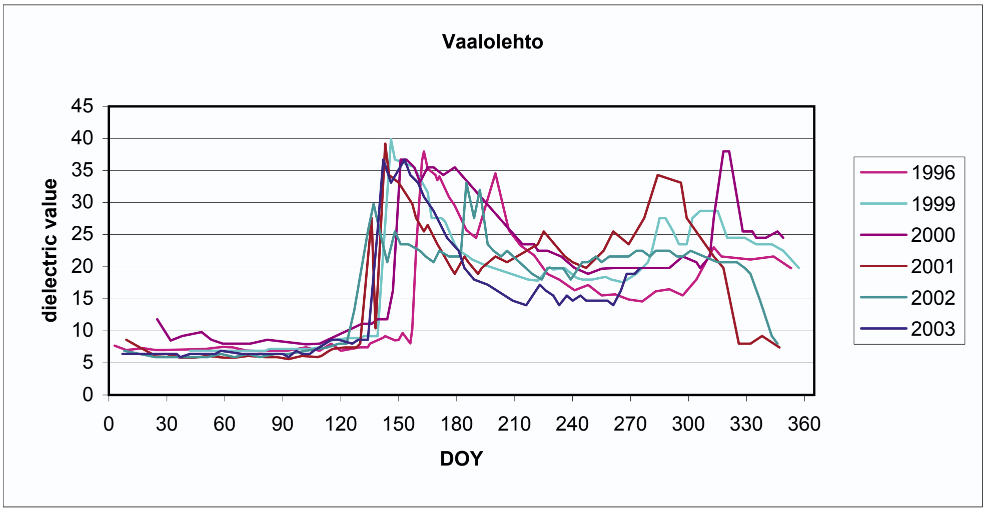


Fig. 5. Annual variation of the dielectric values of silty till at Vaalolehto site presented separately for the years 1995-2003. Note difference between the onsets of water percolation through frozen soil (soil temperature <0ºC) in springs 1996 (June) and 2002 (May).

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THE URBAN GEOLOGY OF TARTU, ESTONIA

by
Volli Kalm[1]

Kalm, V. 2007. The urban geology of Tartu, Estonia. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 141–146, 3 figures.

The territory of the City of Tartu is affected by different natural and human-induced urban geological phenomena, which have been studied since 2003. Subsidence of soils and buildings, strong fluctuations in the groundwater level, contamination of the cultural layer and soils and other undesirable processes are mostly restricted to the oldest city quarters, located on the floodplain of the Suur-Emajõgi river. The highest rate of subsidence of buildings located on the floodplain occurred at the time of the most intensive groundwater consumption from the Quaternary aquifer. Natural deposits of the city territory are covered by a 0.1-12 m thick (mainly 2-3 m) cultural layer that in the city centre is heavily contaminated with lead, zinc and copper.

Key words (GeoRef Thesaurus, AGI): urban geology, ground water, drawdown, land subsidence, soil pollution, heavy metals, Tartu, Estonia.

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INTRODUCTION

The City of Tartu is located in southeastern Estonia (58˚22´N, 26˚43´E) at the river of Suur-Emajõgi, where the river valley (floodplain 32-40 m a.s.l.) between the Devonian sandstone plateaus (50-80 m a.s.l.) is the narrowest (Fig. 1). Tartu was first mentioned in historical records in 1,030 AD, but archaeological finds prove that the first settlements were present between 3-2 millenium BC (Tvauri 2001). The City of Tartu is partly located on the Devonian sandstone plateau while its historical regions lay on soft soils like alluvial, limnic, or bog sediments. The City of Tartu is facing a complex of urban geology problems related to subsidence of soils and buildings, human-induced fluctuations in the groundwater level and distribution of the polluted but archaeologically rich cultural layer, flooding, and slope processes. In 2003, the Tartu Urban Geology Project was initiated and an urban geology GIS-based databank includes information from ca 1,850 boreholes. Some aspects of the ongoing research are published (Karro et al. 2004) or presented in BSc theses (Lokotar 2005, Nõges 2006). This article reports the last observations and interpretations related to the urban geological situation in the City of Tartu.

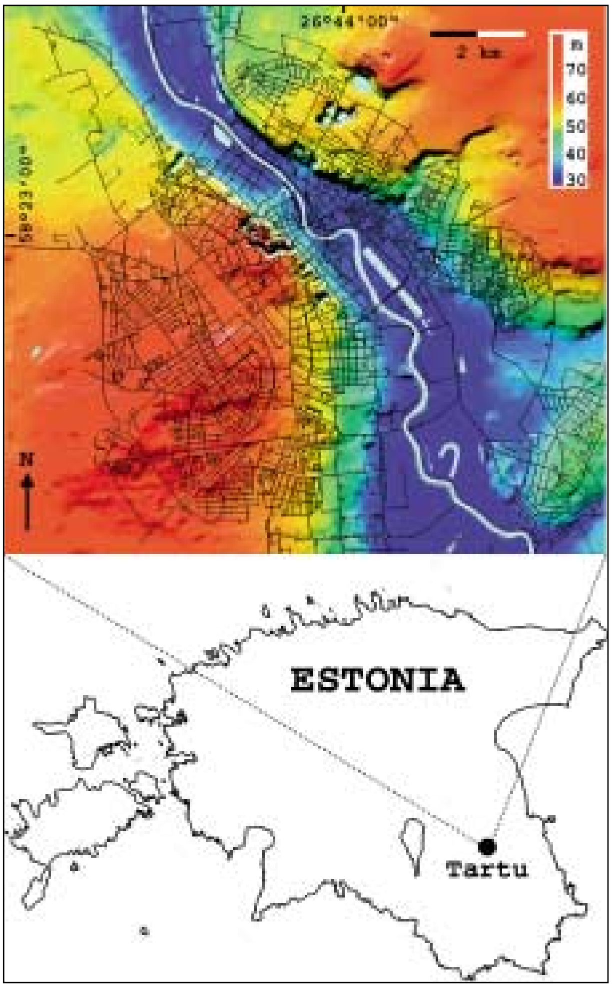

Fig. 1. Topography of the City of Tartu located at the middle reaches of the Suur-Emajõgi River, southeastern Estonia.

GEOLOGICAL BACKGROUND

The Tartu region was deglaciated ca 14.2 kyr BP, after the Otepää Stade and well before the Pandivere (Neva) Stade (Kalm 2006). Under the Holocene alluvial and bog deposits, late glacial varved clays are distributed in the river valley at an elevation up to 26-27 m a.s.l., which indicates that a high-level (>45-50 m) limnic body of water filled the whole valley. On the floodplain area, from top to down, there are more or less continuously present sedimentary units: a) anthropogenic deposits (cultural layer) – 0.1-12 m; b) lake- and spring marl – 0-3 m; c) peat and gyttja – 0-6 m; d) alluvium – 0-2.5 m; e) glaciolacustrine clayey silt and silt – 0-10 m; f) glaciofluvial sand and gravel 0-30 m; g) till – 0-35 m; h) Devonian sandstone. Most of the Holocene deposits (peat, gyttja, spring marl) are weak soils in terms of geotechnical properties. On the Devonian sandstone plateau, geotechnical conditions are good for construction (compressive strength of the sandstone = 6.95 MPa), even though the clayey till, which covers the sandstone, has a higher than average fluidity (yield point 15.95%).

GROUNDWATER LEVEL FLUCTUATIONS AND SUBSIDENCE OF BUILDINGS

Groundwater level fluctuations were reconstructed for the old part of the city based on the data from 114 drillholes made between 1963-2002 (Karro et al. 2004). The monitored area (floodplain) lies mostly between 32-40 m a.s.l., while the average water level in the Emajõgi River (in 1947-1995) next to the monitored area was 30.7 m a.s.l. In 1963, the Toomeoru water intake started to operate some 400-800 m south-southeast of the city centre with five 50-60 m deep wells. The amount of groundwater withdrawn from the Quaternary aquifer varied from 200 to 3,750 m^3/d between 1963-2000 (Fig. 2). Due to intensive consumption, the groundwater level in the water intake lowered 3.5 m by 1974-75 and the groundwater level in the old city area followed that trend (Karro et al. 2004). The lowering of the groundwater level lead to the decrease in pore pressure in unconsolidated soils and to land subsidence (Poland 1985, Chen et al. 2003), which in the Tartu old city area was the highest in 1973 – average of 38 buildings = 4.87 mm/yr (Lokotar 2005). A decrease in groundwater extraction started in 1976 and over the next 20 years, the groundwater level has steadily recovered in the old city area (Fig. 2). According to data from the last general levelling of buildings in 1993, the subsidence has decreased to a rate of 1.5-2 mm/yr (Lokotar 2005).

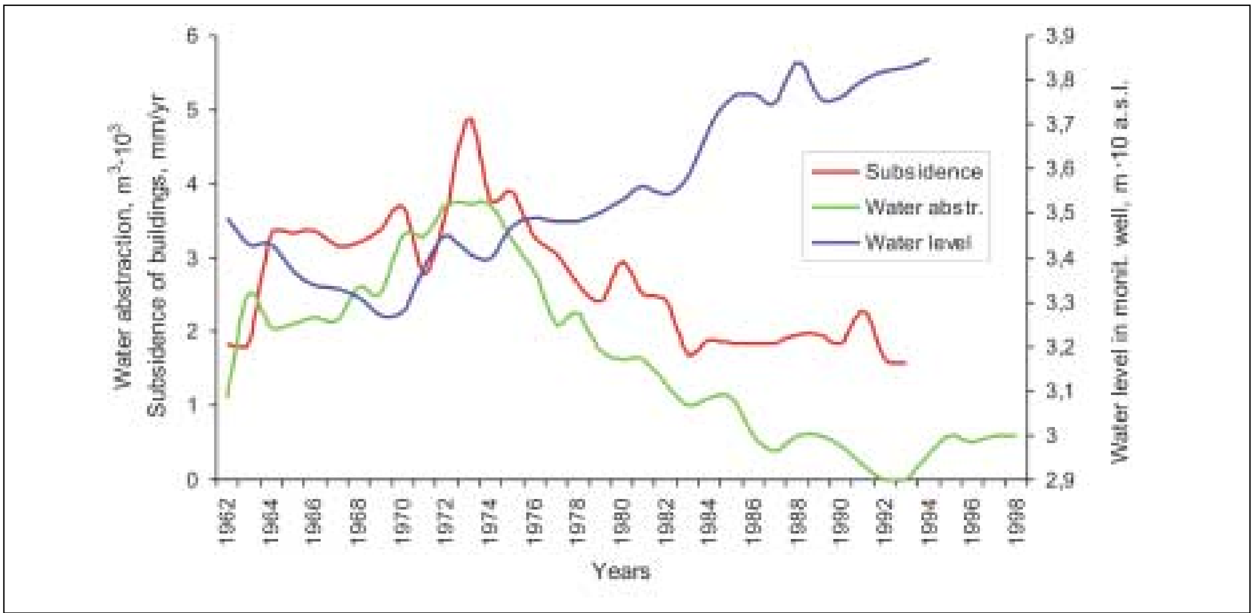

Fig. 2. Average subsidence of buildings located on the floodplain of the Suur-Emajõgi River in relation to water extraction from the Toomeoru water intake and the level of groundwater in monitored wells.

DECAY OF TIMBER IN THE CULTURAL LAYER

Since the 13th century, spruce and pine timber have been placed on peaty soils to support foundations of buildings constructed on peaty soils. Today the timber pavements are ca 2-3 m below the surface under the cultural layer. Due to intensive water extraction and the groundwater level lowering described above, ancient timber pile foundations and rostferk became exposed to aeration. Soon the damage to timber piles by *Cossonus parallelepipedus* worms was determined (Oll 1967) and 14 different species of fungi were detected from the wooden rostferk, which was situated above the groundwater level (Oll 1976). While the foundations of a few historical buildings have been reinforced since 1966, it has not been possible to estimate the contribution of the decay of timber into the rate of total subsidence of the buildings.

HEAVY METALS IN THE CULTURAL LAYER

Chemical analyses of the samples from the cultural layer exposed in an excavation at the city centre revealed that the soil contamination from heavy metals is very high (Fig. 3). The highest concentrations of Pb were found below the layer of debris from the Second World War (WW2), although this probably represents the extensive use of leaded gas after WW2. The maximum estimated concentration of Pb (240 mg/kg; Nõges 2006) in the section is 14.6 times higher than the average amount of Pb in Estonian humic soils (16.4 mg/kg; Petersell et al. 1997). Zinc and Cu reach their maximums (150 mg/kg and 96 mg/kg respectively) in the lower half of the cultural layer, some 1.8-2.6 m below the surface. Concentration values of Zn and Cu have a good correlation (r = 0.77 and 0.82 respectively) with the content of Ca, supporting the conclusion that in Ca-rich soils Cu and Zn may be sorbed on the surface of $CaCO^3$ (Dudley et al. 1991). Concentration of Cd in the cultural layer (average 0.13 mg/kg) remains below the average Cd content in Estonian humic soils (0.4 mg/kg; Petersell et al. 1997). In the peat deposits below the cultural layer, the concentration of heavy metals decreases rather strongly (Fig. 3) and remains close to the average values in Estonian soils.

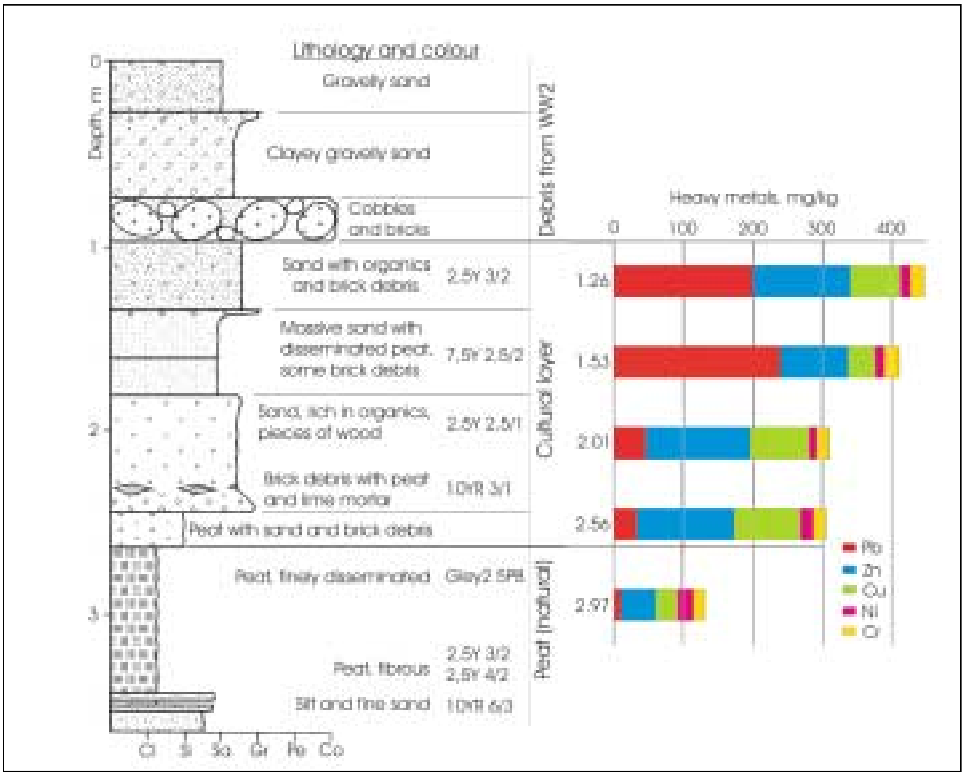

Fig. 3. Heavy metals distribution in the section through cultural layer in the centre (corner of Küüni and Vanemuise streets) of the City of Tartu.

CONCLUSIONS

Undesirable urban geological phenomena – subsidence of buildings, groundwater level fluctuations, soil pollution – are mostly restricted to the city quarters located on the floodplain area. Lowering of the groundwater level in the floodplain deposits lead into soil compaction, decay of timber rafts under the foundations of buildings and subsidence of buildings. The highest rate of subsidence (4.9 mm/yr in 1973) was recorded at the time of the most intensive groundwater consumption from the Quaternary aquifer. Since 1974, the groundwater level has risen and the subsidence decreased to 1.5-2 mm/yr over a 20 year period. Initial analyses showed that the cultural layer in the centre of the city is heavily contaminated with Pb, Zn and Cu, but not with Cd.

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SEDIMENTOLOGICAL CHARACTERISTICS OF LATE WEICHSELIAN–HOLOCENE DEPOSITS OF THE SUURPELTO AREA IN ESPOO, SOUTHERN FINLAND

by
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Ojala. A. E. K. & Palmu, J.-P. 2007. Sedimentological characteristics of Late Weichselian–Holocene deposits of the Suurpelto area in Espoo, southern Finland. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 147–156, 8 figures, 2 tables.

Lithostratigraphic studies and analyses of the physical characteristics of sediments have provided a detailed picture of the fine-grained deposits in the Suurpelto area in Espoo, southern Finland. A key sediment section from drill-hole N5, which represents different depositional phases of the Baltic Sea Basin (BSB) history, was divided into 11 lithostratigraphic units according to their sediment structure and composition. The properties of these units were carefully described. A 3D geological model of these units in the Suurpelto area has been established, including cross-section profiles from northern and southern parts of the basin. The northern part of the Suurpelto basin contains a thinner sequence of massive and varved clays with coarser (silt, sand) mineral layers representing earlier phases of the BSB than the southern part of the basin. In the southern part of the basin these deposits have been covered by more organic and sulphide-rich sediments that represent later phases of the BSB. Three locations in the Suurpelto area have thicker than average (10.6 m) sequence of fine-grained deposits. The maximum thickness of fine-grained sediment in the Suurpelto area is >25 m in the SW corner.

Key words (GeoRef Thesaurus, AGI): sediments, clay, lithostratigraphy, physical properties, thickness, Holocene, Pleistocene, Weichselian, Suurpelto, Espoo, Finland.

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INTRODUCTION

Reconstruction of the geological history of the Suurpelto area is an important task because the city of Espoo has recently begun land-use planning of the area. The area has deposits of silt and clay up to 25 m thick. These fine-grained sediments in the Suurpelto area have accumulated since the Fennoscandian ice-sheet retreated from the area ca. 13,000 years ago, during the early history of the Baltic Sea Basin (BSB) (e.g. Björck 1995, Saarnisto & Saarinen 2001). Based on geotechnical information it is known that the composition and geotechnical properties of the Suurpelto deposits vary both vertically and horizontally, making it a challenging site for construction. Therefore, the Suurpelto project was launched in 2005 in cooperation with the city of Espoo, the Geological Survey of Finland, the University of Helsinki and Helsinki University of Technology, aiming to provide scientific and technical information for land-use planning and ground engineering, particularly for tasks such as construction suitability and stabilization. Geophysics, geochemistry, geotechnical engineering and sedimentological methods have been applied in the Suurpelto project (Ojala et al. 2006, Palmu et al. 2006). One of the purposes of the project is to compare and combine results gained with different investigation methods, and in the longer term, to develop new methods or method combinations that could be efficiently and economically applied to similar areas.

This paper presents results from the sedimentological part of the Suurpelto project. It includes the methods of logging several physical properties of fine-grained sediment deposits and a highly detailed characterization of sediment lithostratigraphy. The primary objectives were to identify the main sedimentary units of the Suurpelto deposits, to characterize their composition and structure, and to construct a 3D geological model of the arrangement of these different units in the Suurpelto area.

STUDY SITE

The Suurpelto area is about 1 km wide and 2 km long and located in Espoo, on the southern coast of Finland, about 10 km west of Helsinki city centre (Figs. 1 and 2). Nowadays the Suurpelto area is situated 4 km from the coastline, lying between 5 and 10 metres above sea level (a.s.l.), and has a current land uplift rate of 2 mm yr^{-1} (Kakkuri 1991). It is surrounded in all directions by a rolling relief of bedrock topography with relative altitude differences of about 50 m. The bedrock is an ancient peneplain of crystalline amphibolites, gneisses and granites that slopes towards the south with a very low gradient (Laitala 1991). Quaternary deposits up to 30 m thick have filled bedrock fracture zones and depressions during glacial and post-glacial times. The basal Quaternary deposits consist of a thin till cover overlain by a glaciofluvial sand deposit of approximately 2 m in thickness in the Suurpelto area. Upon these lies a section of fine-grained sediments, the primary target of the present investigation, that was formed during the retreat of the continental Fennoscandian ice-sheet from the area and during different phases of the BSB deposition.

The geological history of the Suurpelto area has been closely connected with the BSB deposition during Late Weichselian and Holocene. Following deglaciation (ca. 13,000 yrs ago), transgressive waters of the Baltic Ice Lake reached their maximum altitude of 120 m a.s.l. in the Espoo area. Subsequently, the lake suddenly drained westwards to the Atlantic and the water table lowered by 26 m at around 11,600 years ago (Eronen 1990, Saarnisto & Saarinen 2001). The emergence of the Suurpelto area proceeded during the Yoldia Sea phase, with the highest sea level of about 80 m a.s.l. in Espoo area (Eronen 1990). It was followed by Ancylus Lake phase (ca. 10,700-9,000 years ago) and Litorina Sea phase (9,000-) of the BSB history. Exposure of the Suurpelto study area from underneath the regressive Litorina Sea waters began around 3,000 years ago (Hyvärinen 1999). An isolation threshold of the Suurpelto basin is situated in the SE corner, nowadays named the Lukupuro ditch, at an altitude of 6 m (Fig. 2). The Suurpelto area became isolated from the Litorina Sea of the BSB around 2,000 years ago (Hyvärinen 1999) and then gradually dried up.

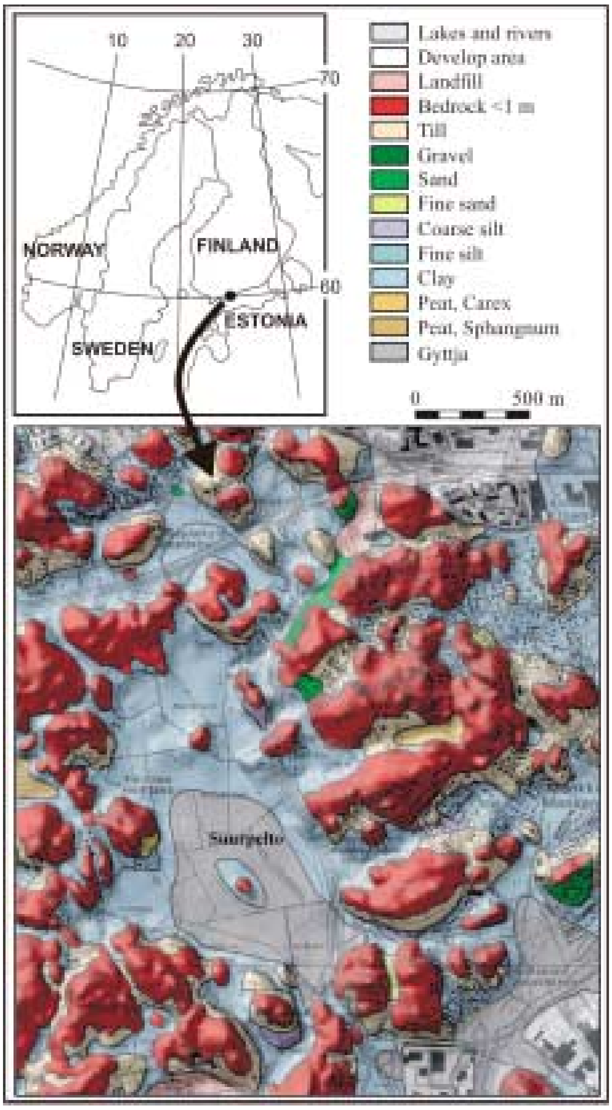

Fig. 1. Location and Quaternary deposits of the Suurpelto area. Digital elevation model and basemap © National Land Survey of Finland, permit No. 13/MYY/07.

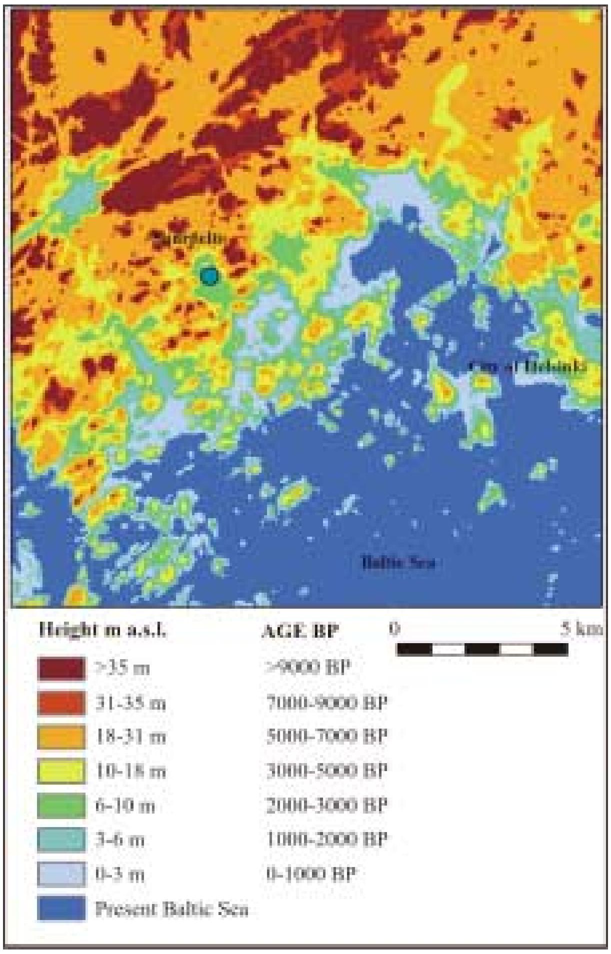

Fig. 2. Shoreline displacement in southern Finland during the last 9000 years according to Hyvärinen (1999). The study site, Suurpelto, is marked with a blue circle. Digital elevation model and basemap © National Land Survey of Finland, permit No. 13/MYY/07.

It is important to note that during deglaciation and the different phases of the BSB the sedimentary environment in the Suurpelto area varied significantly due to several reasons. The primary variables were the discharge intensity of melt-water from the retreating Fennoscandian ice-sheet, the relative water depth (sea level rise vs. glacioeustacy), the water salinity (i.e. fresh water/brackish water), and the intensity of primary productivity due to climate change and nutrient availability. Therefore, the composition and structure of the Suurpelto fine-grained deposits has varied through time and space.

MATERIALS AND METHODS

Drilling and lithostratigraphic logging

The results of this study are based on three sources of information (Fig. 3). Firstly, six drill-holes (N1-N6) that were taken with a piston-type lake sediment corer (Putkinen & Saarelainen 1998) modified for the purpose of retrieving continuous sections of clayey deposits into ca. 1-metre-long plastic tubes (sample tube inner diameter of 45 mm). Core samples were taken in 2005-2006. Secondly, 26 coring locations logged with a small (diameter 50 mm, length 500 mm) Russian peat corer operated with a series of 1-metre-long steel rods in 2006. Thirdly, results from 311 Swedish weight-sounding tests carried out by the city of Espoo between 1968 and 2006. The last two information sources were used to characterize the sediment lithostratigraphy and thickness of fine-grained deposits in the Suurpelto area, whereas the piston core samples were analysed in the laboratory to determine the physical and chemical properties of the sediment.

Sediment lithostratigraphy and physical analyses

The sample tubes of piston cores N1-N6 were opened in the sediment laboratory by splitting them lengthwise to enable sediment description and analyses. Sections of up to 21 m long, the longest being drill-hole N5, were described in detail according to the classification of Troels-Smith (Troels-Smith 1955). Magnetic susceptibility (MS) was then measured at 2 and 1 cm resolution using a Bartington MS2C loop sensor and MS2E1 surface scanning sensor, respectively. MS measures the concentration and grain-size of magnetic minerals in deposits, and can also be used in correlation between parallel cores (e.g. Sandgren & Snowball 2001). Water content (WC per dry weight) and loss on ignition (LOI) were then determined at ca. 10 cm resolution by drying the samples overnight (105 ºC) and burning in a furnace (550 ºC) for 2 hours. Dry density was determined according to the formula by Håkanson and Jansson (1983). Additional mineral magnetic parameters, namely anhysteretic remanent magnetization (ARM), saturation isothermal remanent magnetization (SIRM), the S-parameter (determined as IRM $-_{100\,\mathrm{mT}}$/IRM $+_{1000\,\mathrm{mT}}$) and their ratios, were applied at ca. 10 cm resolution to investigate variations in the quality and concentration of magnetic minerals (e.g. Sandgren & Snowball 2001).

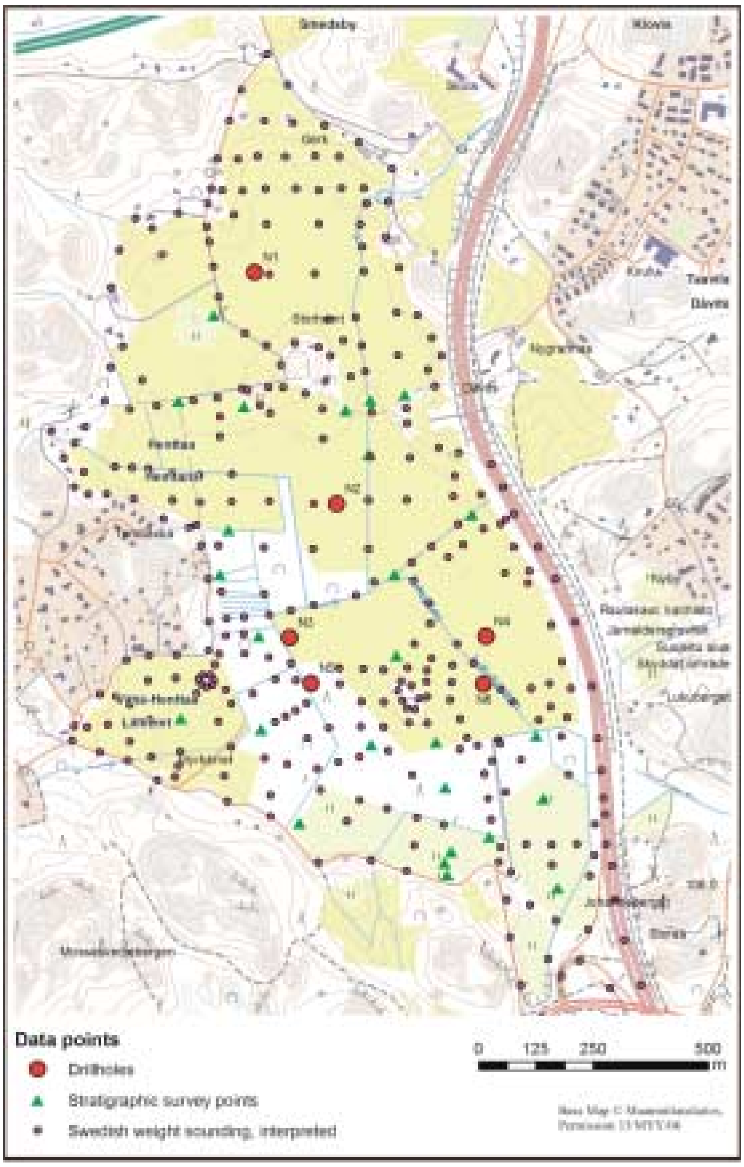

Fig. 3. Locations of drill-holes N1-N6 (red dots), stratigraphic survey points (green triangles) and Swedish weight-sounding tests (purple dots) used in the present study. Digital elevation model and basemap © National Land Survey of Finland, permit No. 13/MYY/07.

RESULTS AND DISCUSSION

Sediment lithostratigraphy and physical properties

One example of the sediment lithostratigraphy in the Suurpelto area from drill-hole N5 is illustrated in Fig. 4. Being one of the earliest places in Finland exposed from beneath the continental ice-sheet, the thickest sections of the Suurpelto fine-grained sediment deposits show very nicely the lithostratigraphic sequence of BSB history. Based on sedimentological analysis, 11 sedimentary units differing in composition and structure were identified from the 21-metres-long section of drill-hole N5. These represent deposits ranging from clays deposited in Baltic Ice Lake to the dried up lake/swamp gyttja and peat from the late depositional phase of the basin (Fig. 4). Their Troels-Smith classification details are presented in Table 1.

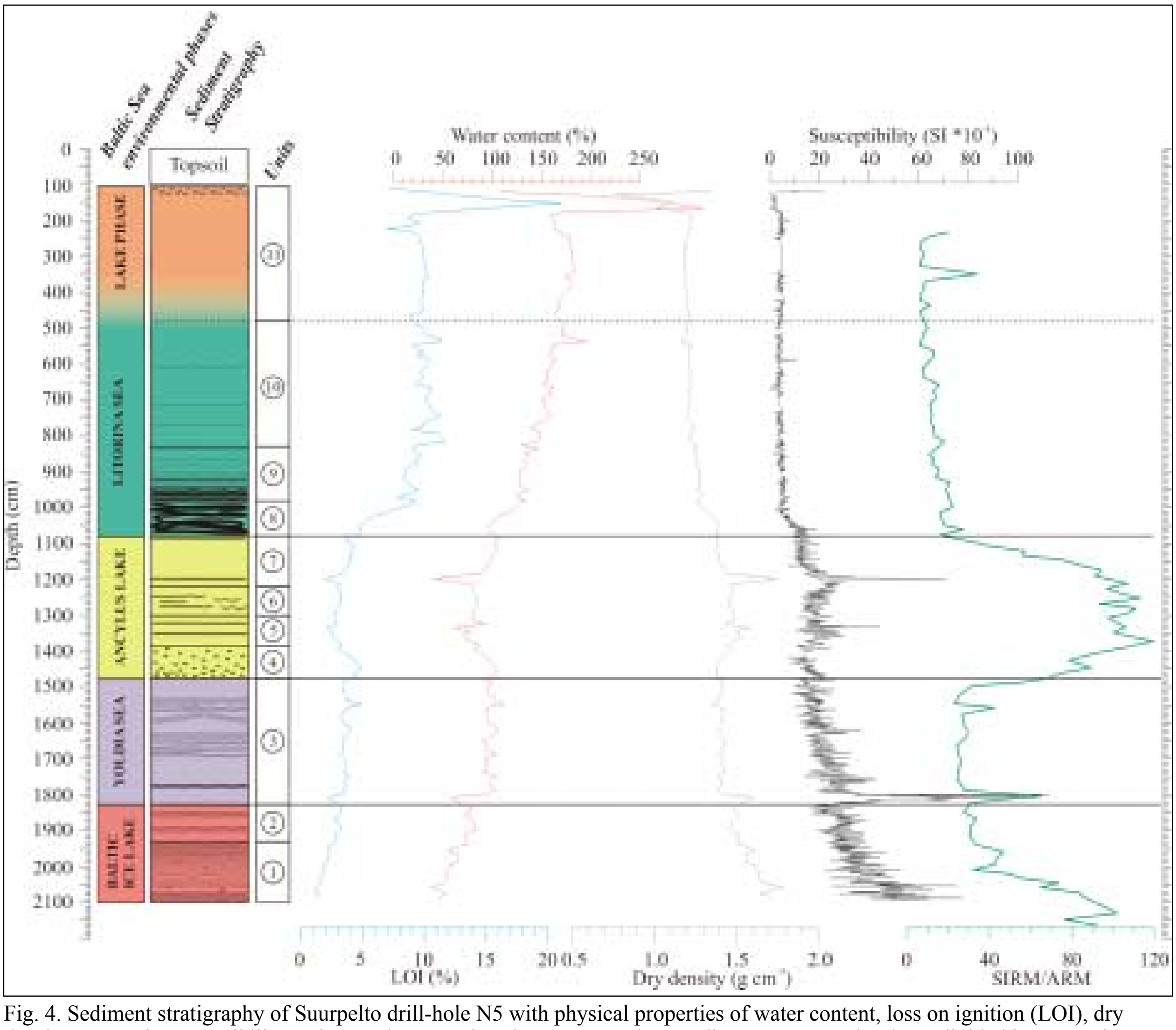

Fig. 4. Sediment stratigraphy of Suurpelto drill-hole N5 with physical properties of water content, loss on ignition (LOI), dry density, magnetic susceptibility and SIRM/ARM ratio. The 21-metres-long sediment sequence has been divided into 11 units according to its sediment structure and composition. Troels-Smith classification codes of these units are presented in Table 1.

The lowermost units 1 and 2 represent deposits sedimented in Baltic Ice Lake and are mainly composed of varved clays (see e.g. Sauramo 1923) with coarser 1-20 mm thick sub-layers of silt and sand, and occasional drop-stones. These units have a very low LOI (< 2%) and high MS and dry density values of up to 80 SI $\times 10^{-5}$ and 1.4 g cm^{-3}, respectively. Their deposition represents the period when the Fennoscandian ice-sheet retreated from the area. The varved structure reflects the annual cycle of sedimentation (e.g. Sauramo 1923). As a rule of thumb, the coarsest layers and thickest varves are found in the lower part of Baltic Ice Lake deposits when the ice was closest to the study area. The varved structure was also controlled by the intensity of spring and summer melt-water discharge from the ice-sheet.

Unit 3 probably represents deposits sedimented in Yoldia Sea when the Baltic basin was temporarily connected to the Atlantic, allowing saline waters to spread into the ancient BSB. It starts with a massive greyish 30 cm thick layer of clay that is very different from the underlying reddish-coloured Baltic Ice Lake varved clays. This layer is easily identified from the MS curve with anomalously high values and possibly indicates the rapid sea level fall that ended the Baltic Ice Lake phase at around 11,600 years ago. In general, unit 3 is characterized by a stable LOI and WC, and as discussed by Gardemeister (1975), sediments deposited in Yoldia Sea phase are typically very rich in clayey (< 0.002 mm Ø) mineral matter with a content of up to 80%. A high clayey contents was observed in here in geotechnical investigations (Ojala & Tanska 2006). Unit 3 contains sections with layered sediment structures that are partly disturbed.

Units 4 to 7 include fine-grained sediment deposits that accumulated during the Ancylus Lake phase of BSB history. At that time the Baltic basin was a freshwater lake that was first transgressive with a rising and then regressive with a lowering water level (Björck 1995). The boundary between units 3 and 4 is very distinctive. Unit 4 consists of massive sulphide-bearing clay that begins with sulphide grains of up to 4 mm in diameter frequently occurring within clay-rich deposits. The quantity of these grains declines upwards in this unit and there are only scattered remains in the overlying unit 5. Unit 5 consist of greenish grey clay containing 2-5 mm thick sub-layers of silt and sand, possibly related to Ancylus Lake water level fluctuations. Units 6 and 7 again contain sulphide as laminae, but not as high in content as in unit 4. The physical characteristics of the sediment indicate that units 4-7 have a lower LOI and WC than the under- and overlying deposits, which according to Gardemeister (1975) is very typical for Lake Ancylus sediments. The deposits sedimented in Lake Ancylus phase is also very well distinguished in the magnetic mineralogy, particularly in susceptibility and SIRM/ARM ratio plots (Fig. 4). This section has a greater magnetite concentration and/or decreased grain-sizes.

Units 8 to 10 represent the Litorina Sea phase of BSB history. These units have a substantially higher WC and LOI than the underlying BSB phases. WC and LOI increase upwards in the Litorina Sea

Table 1. Troels-Smith classification codes (Troels-Smith 1955) for the Suurpelto sedimentary units.

Unit	Troels-Smith description
11	nig. 2, strfc. 0, elas. 0, sicc. 1 colour: brown struc: massive clay gyttja comp: As 3, Ld 1
10	nig. 2, strfc. 1, elas. 0, sicc. 1 colour: dark brown struc: sulphide laminated gyttja clay comp: As 3, Ld 1, sulf +
9	nig. 3, strfc. 3, elas. 0, sicc. 1 colour: black/grey struc: sulphide striped gyttja clay comp: As 3, Ld 1, sulf +
8	nig. 4, strfc. 1, elas. 0, sicc. 1 colour: black struc: sulphide clay comp: As 3, Ld1 sulf ++
7	nig. 2, strfc. 0, elas. 0, sicc. 1 colour: dark grey struc: clay with sand laminae comp: As 4, Gmin+, sulf+
6	nig. 2, strfc. 1, elas. 0, sicc. 1 struc: sulphide laminated clay comp: As 4, sulf+
5	nig. 2, strfc. 0, elas. 0, sicc. 1 colour: grey struc: massive clay with sand laminae comp: As 4, Gmin+
4	nig. 2, strfc. 0, elas. 0, sicc. 1 colour: dark grey struc: massive clay with suplhide grains comp: As 4, sulf+
3	nig. 1, strfc. 2, elas. 0, sicc. 1 colour: grey struc: partly laminated clay comp: As 4
2	nig. 1, strfc. 2, elas. 1, sicc. 1 colour: reddish grey struc: partly laminated clay comp: As 4, Ag+
1	nig. 1, strfc. 4, elas. 0, sicc. 1 colour: grey struc: silt/sand laminated clay comp: As 4, Ag+, Gmin+

deposit from 10 to 20% and 5 to 12%, respectively. In addition, susceptibility rapidly decreases and then remains very low (< 5 SI 10^{-5}) and stable. Unit 8 is very rich in sulphide and has a distinctive lower boundary with the underlying unit 7. The black-coloured sulphide layer is more than a metre thick and then grades upwards into sulphide-laminated gyttja clay/clay gyttja in unit 9. Unit 10 consists of massive clay gyttja that represents a period prior to the isolation of the Suurpelto area. By 3,000 BP the Suurpelto area had become a sheltered gulf of the BSB, probably with a fairly calm sedimentation environment and a water depth of about 10 metres (see Fig. 2). The area is surrounded by a rolling relief of bedrock topography (with an altitude of 30 to 50 m a.s.l.) in all directions, and has therefore been well sheltered from the wind and major wave activity. It is likely that a lake existed in the Suurpelto area after isolation from the Litorina Sea at around 2,000 years ago, which then gradually dried up. Organic gyttja deposits in the upper part of the section (unit 11) were deposited during this period. Its lower boundary is gradual, and it is therefore impossible to identify the exact isolation horizon in the sediment lithostratigraphy. Above unit 11 lies a ca. 1-metre-thick section that is under constant change due to soil formation and agricultural activity.

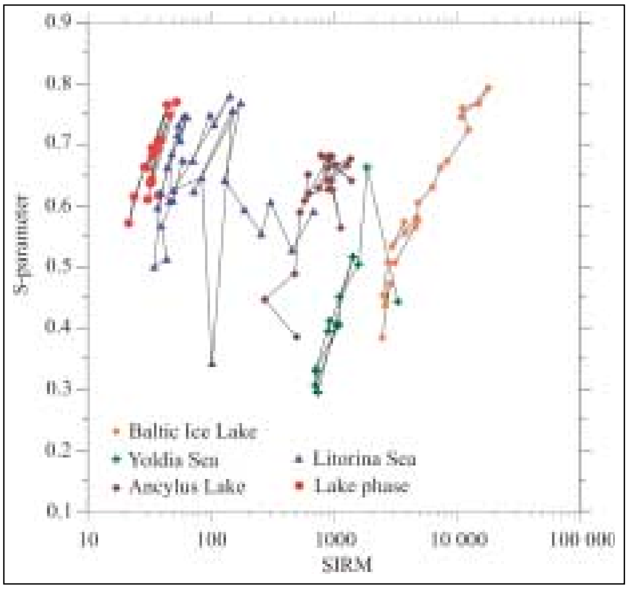

Fig. 5. Magnetic information, in this case SIRM vs. S-parameter, can be applied to identify different sedimentological units according to their quality, concentration and the grain-size of magnetic minerals. Here, the deposition of different stages of geological history of the Suurpelto basin appear from the far right (Baltic Ice Lake deposits) towards the far left (Lake phase deposits). For parameters see Sandgren and Snowball (2001).

The above-described sediment lithostratigraphical division is well represented in the SIRM vs. S-parameter plot of the sediment magnetic properties (Fig. 5). The characteristics of this figure mainly reflect fluctuations in magnetite grain-size and concentration, seemingly allowing a clear distinction of the deposition of the different sedimentation environments in the Suurpelto basin.

Sediment distribution in the Suurpelto area

The same physical units that represent different phases of BSB history in drill-hole N5 (Fig. 4) are found almost throughout the Suurpelto area. However, due to variations in basin topography, water depth and the location of the isolation threshold, their distribution and thickness vary in the area. In addition, changes in wave activity and water currents during and prior to isolation have caused substantial erosion and redeposition of material in the Suurpelto basin. One of the primary aims of this project was to establish a 3D geological model of the sedimentological units within the Suurpelto basin. We aimed to improve our general understanding of these fine-grained deposits in order to be able to better compare their geological properties with engineering properties in different parts of the Suurpelto area.

Fig. 6 illustrates the thickness and distribution of different section type of deposits of the fine-grained sediments, grouped as Topsoil, deposits of the Lake phase (unit 11), deposits of the Litorina Sea (units 10, 9. 8), deposits of the Ancylus Lake (units 7, 6, 5, 4), deposits of the Yoldia Sea (unit 3) and deposits of the Baltic Ice Lake (units 2, 1) in the Suurpelto area. These sections have been combined in Fig. 7, which characterizes the overall thickness variability of fine-grained sediments in the Suurpelto area. From top to bottom, the thickness of the topsoil in the Suurpelto area is less than a metre thick, but mostly 60 to 80 cm. The composition of this topsoil varies between sites but is generally either organic peat or gyttja, or agricultural soil that is under constant change.

The deposits of the Lake phase are mostly distributed in the lowermost parts of the Suurpelto area in the south, i.e. places where the ancient lake or swamp was located after isolation. The thickest Lake phase sections, up to 4 m in thickness, are in the SW sector of the basin. Likewise, Litorina Sea deposits are thickest in the SW part of the Suurpelto area, being up to 8 m thick. This is important because the lower part of Litorina Sea deposits is richest in sulphide, which may affect construction suitability and stabilization. As seen in Fig. 8, the sulphide-rich horizon lie at a depth of about 10 and 7 m below the sediment surface (i.e. 4 and 1 metres below the present sea level) in the SW and SE depressions, respectively. There is only a very thin or nonexistent layer of organic sediments (from the Lake phase and Litorina Sea phase) in the northern part of the Suurpelto basin. The fringes and south side of the exposed bedrock in the middle of the southern Suurpelto basin also lack organic deposits. These areas lie at a higher elevation and/or have been sensitive to erosion during the isolation of the Suurpelto area. It is likely that eroded material has been redeposited into basin depressions, thus increasing their rate of accumulation.

Sediment of the Lake Ancylus generally follow the same pattern of distribution as the more organic overlying sediments. The thickest sections, up to 6 m thick, are found in the southern part of the Suurpelto area. These sections also contain sulphide clays that lie on average 3-4 m below the Litorina Sea sulphides. The summed sediment volume of units 7, 6 and 5 is of a similar magnitude to that of the summed units 10, 9 and 8 (Table 2).

Deposits of the lowermost three units (units 1, 2 and 3) are then more widely distributed throughout the Suurpelto area. In particular, varved and massive clays of Baltic Ice Lake form a thick lowermost section of the fine-grained deposits of Suur-

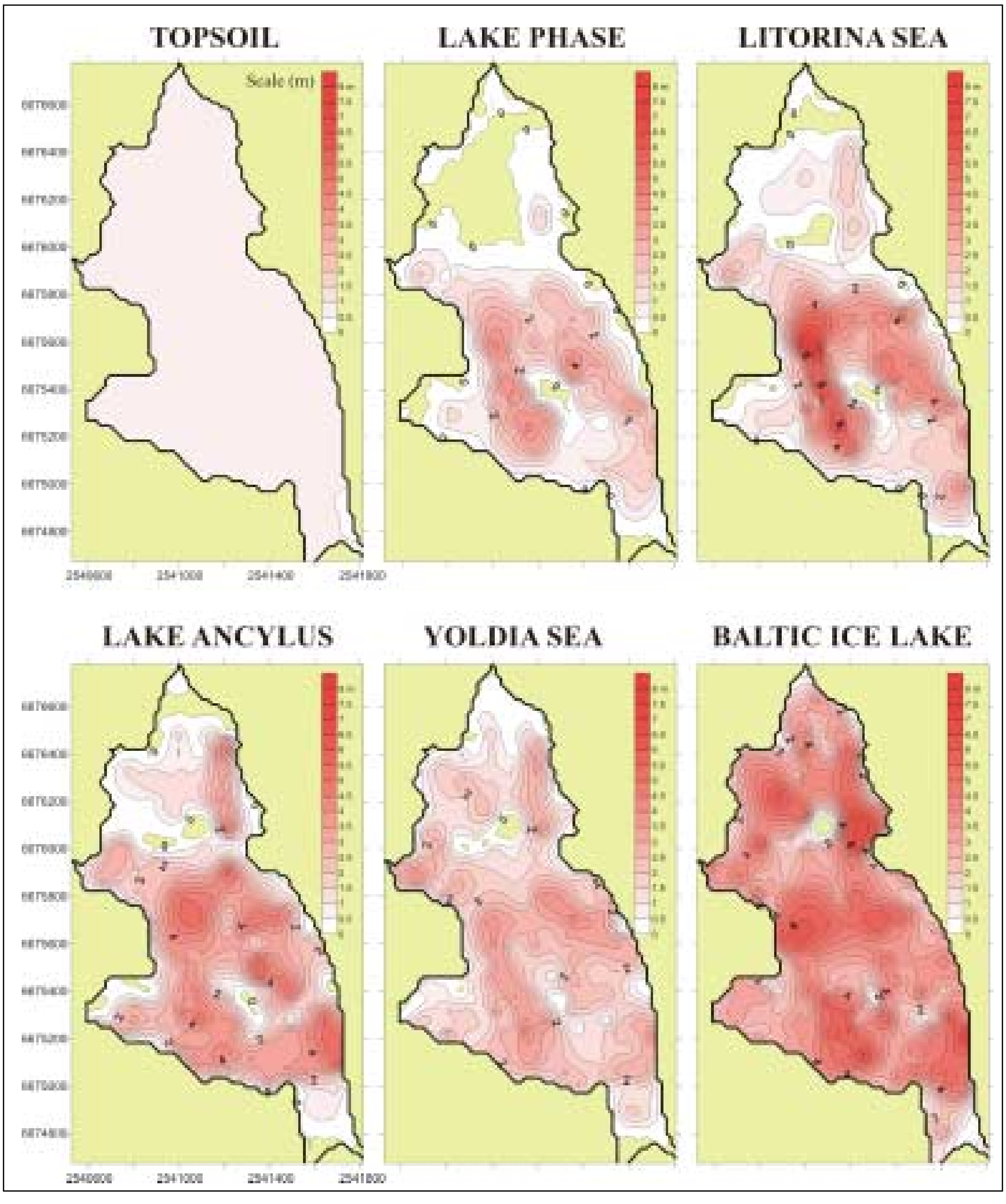

Fig. 6. Thickness and distribution of sediments in the Suurpelto area deposited in different Baltic Sea phases. Note that the most organic deposits (from the Lake and Litorina Sea phases) are located in the southern part of the basin.

pelto that extends all around the basin, including the fringe areas. Deposits of varved and massive clayey lie directly below the topsoil in the fringe areas of the Suurpelto basin. Their volume is about twice as great as that of other BSB phases and the deposits are mostly evenly distributed.

The pattern of total thickness of fine-grained sediment deposits (Fig. 7) correlates rather well with the thicknesses of the uppermost 4 or 5 sedimentary units. There are three main basins in the Suurpelto area that have thicker than average (10.6 m) sections of fine-grained deposits. These basins follow bedrock fracture zones. The first is located in the SW part of the Suurpelto basin. It is about 200 m wide and 800 m long with a maximum sediment thickness of > 25 m, and is the main sedimentary basin that has effectively accumulated primary and secondary material from erosion. The second basin in the SE part of the Suurpelto area is also elongated in a SE-NW direction. It has a maximum sediment thickness of 22 m. The third basin is located in the NE part Suurpelto with up to 15 m thick sections of fine-grained sediments.

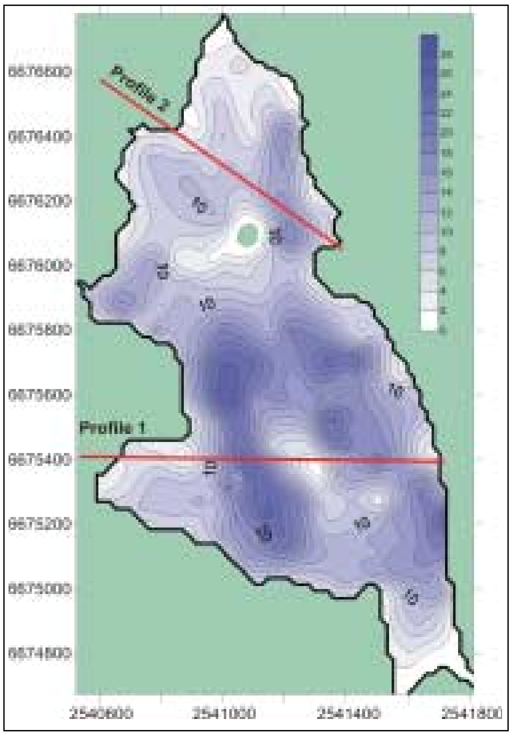

Fig. 7. The overall thickness of fine-grained deposits in the Suurpelto area according to the presently studied locations (see Fig. 3). Red lines indicate the cross-section of profile lines 1 and 2 presented in Fig. 8.

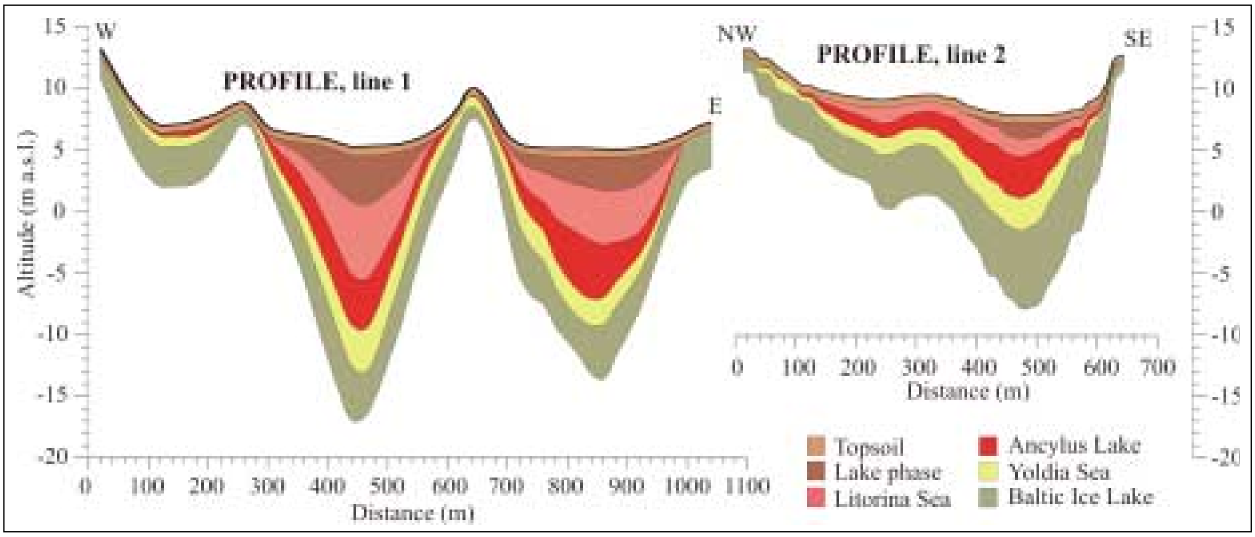

Fig. 8. Cross-sectional profiles from the Suurpelto area (see Figure 7 for the location of these profiles). They indicate that there is considerable variation in the thickness of fine-grained deposits in the Suurpelto are in terms of the overall thickness and the thickness of different units.

Table 2. Sediment volumes and positive surface areas of different phases of the BSB in the Suurpelto area.

Stratigraphic unit	Volume (milj. m^3)	Area (ha)
Deposits of “Top soil”	0.81	125
Deposits of Lake phase	1.2	102
Deposits of Litorina Sea	2	116
Deposits of Ancylus Lake	2.5	118
Deposits of Yoldia Sea	2.2	123
Deposits of Baltic Ice Lake	4.7	126

In summary, the northern part of the Suurpelto area has a thinner sequence of fine-grained deposits. There are mainly massive clays with a varved structure and coarser (silt, sand) mineral layers that represent earlier periods in BSB history, namely the Baltic Ice Lake and Yoldia Sea phases. In the southern part of the Suurpelto area these Baltic Ice Lake and Yoldia Sea deposits have been covered by more organic and sulphide-containing deposits representing later phases of the BSB and an ancient isolated Suurpelto lake phase. The southern part is also characterized by thicker sequences of fine-grained material, partly from primary accumulation and partly from erosion and the redeposition of secondary material.

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MORPHOTECTONIC INVESTIGATIONS OF THE ANTARCTIC

by
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Lastochkin, A. N. & Krotova-Putintseva A. Y. 2007. Morphotectonic investigations of the Antarctic. Applied Quaternary research in the central part of glaciated terrain. *Geological Survey of Finland, Special Paper 46*, 157–161, 3 figures.

New data on the subglacial-submarine relief structure of the Antarctic were obtained during geomorphologic investigations on the basis of the morphodynamic concept and system-morphologic principle. General geomorphologic investigations were followed by special morphodynamic investigations, including morphotectonic interpretations of general geomorphologic and morphometric maps as well as available geological and geophysical data. As a result, a morphotectonic map of the Antarctic is presented.

Key words (GeoRef Thesaurus, AGI): geomorphology, tectonics, morphostructures, tectonic maps, Antarctica.

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INTRODUCTION

Hypsobathymetric data (Lythe et al. 2000) and the morphodynamic concept with its system-morphologic principle (Lastochkin 1991), applied to the subglacial-submarine surface of the Antarctic have provided new notions about the morphotectonics of the Antarctic.

We assumed that the lifetime of ice cover – 25-35 million years (Losev 1982) is commensurable with the neotectonic period of Earth crust evolution. In general, neotectonic movements defined the height and depth amplitude of modern relief, and morphology and contrast of its forms. The latter were preserved from lithodynamic processes and are the most objective indicators of concurrent tectonic movements.

PRINCIPLES OF MORPHOTECTONIC ZONING OF THE ANTARCTIC

A morphotectonic map (Figs. 1, 2, 3) is the result of morphotectonic interpretations of subglacial-submarine relief orography. The morphotectonic map shows morphotectonic regions (morphotectonic

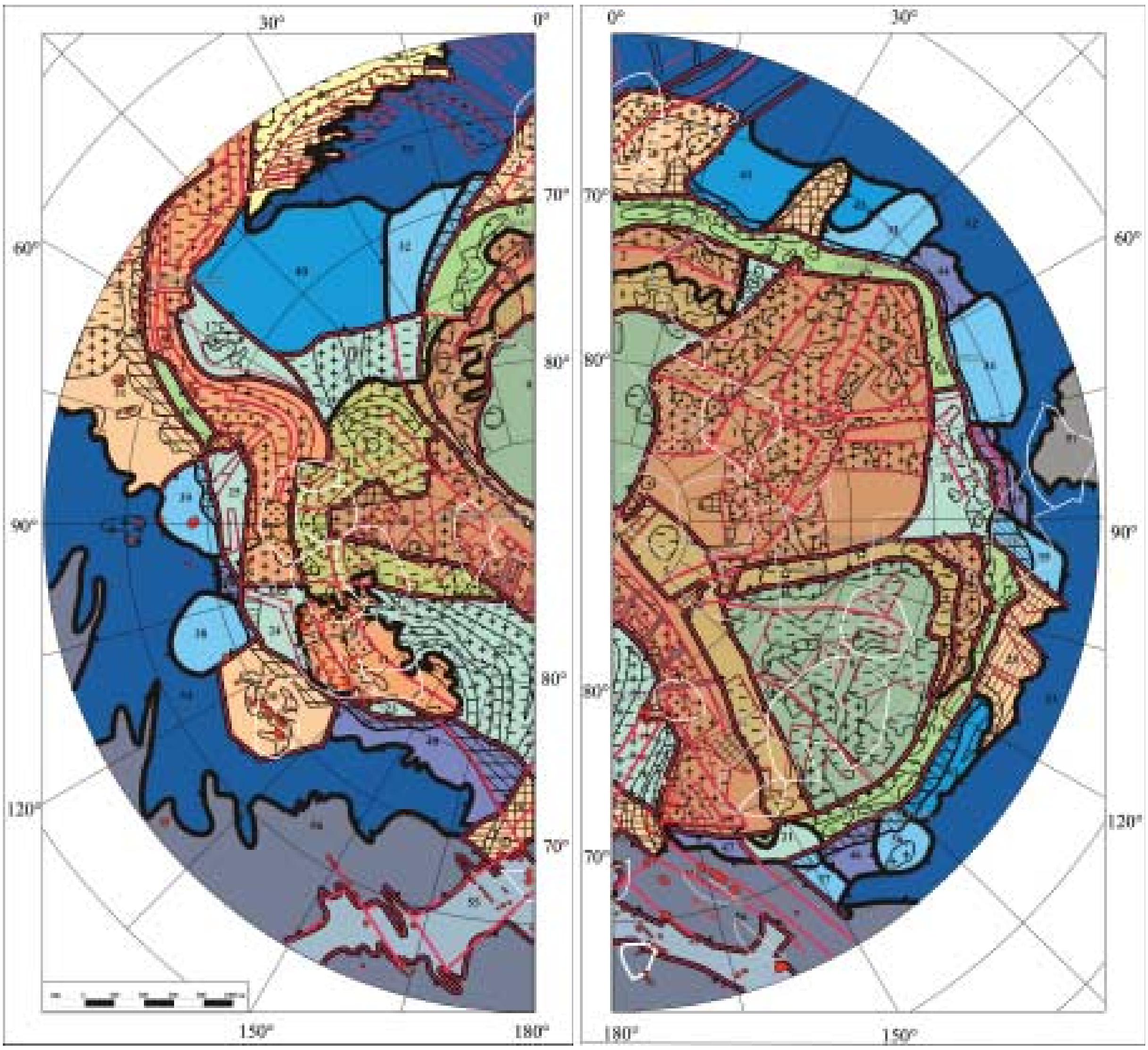

Fig. 1. Morphotectonic map of the Antarctic (authors: Lastochkin A. N., Krotova-Putintseva A. Y.)

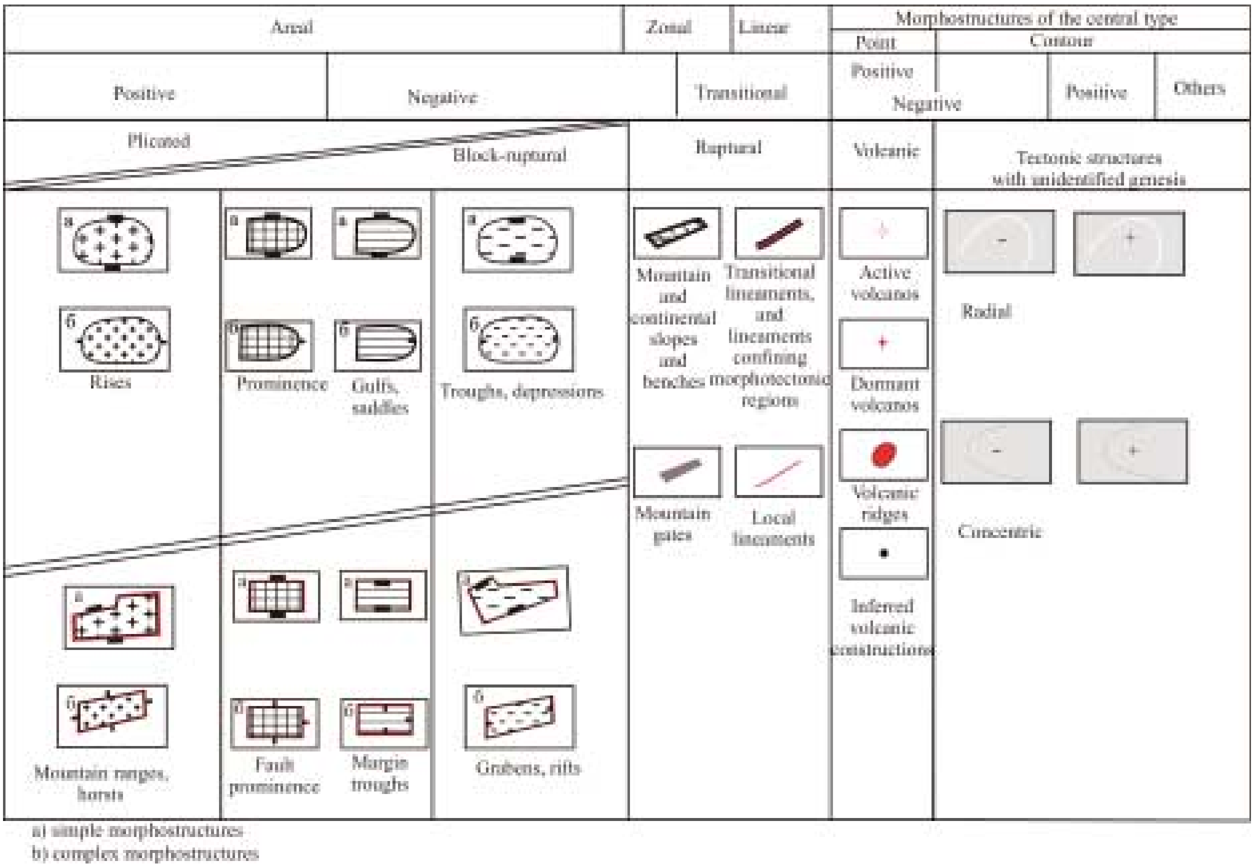

Fig. 2. Legend of the morphotectonic map of the Antarctic (morphostructures).

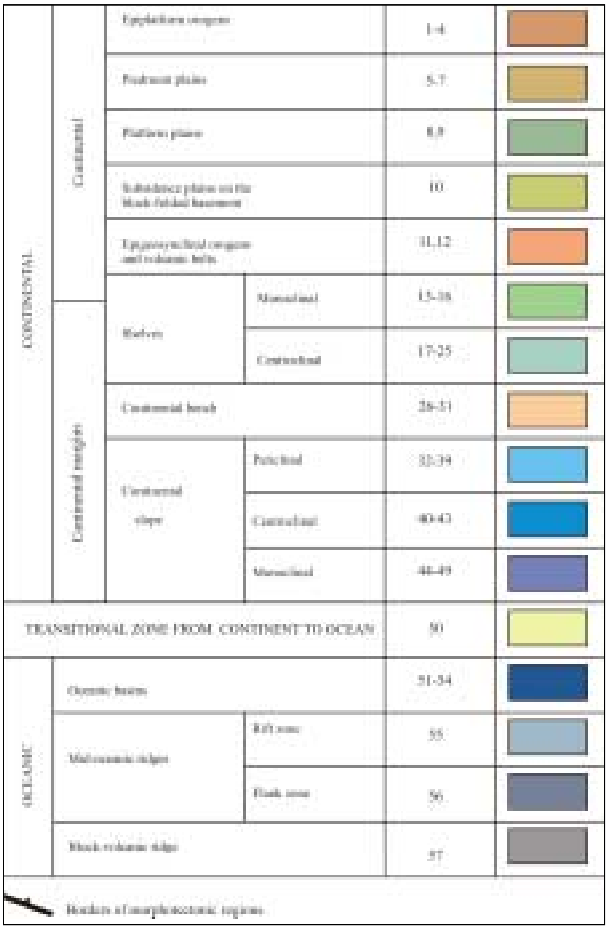

Fig. 3. Legend of the morphotectonic map of the Antarctic (morphotectonic regions).

zoning) and areal, linear and point morphostructures. Areal morphostructure is considered to be a three-dimensional morphogeodynamic formation – a part of the Earth's surface and the corresponding block of the Earth's crust, which as a whole undergoes tectonic and/or isostatic movements of certain intensity and direction. Linear morphostructures (lineaments) imply expression of disjunctive dislocations in the subglacial-submarine relief of the Antarctic. Lineaments confine and intersect with areal morphostructures. Point morphostructures include active and dormant volcanoes revealed on geological and geophysical data (Gonzalez-Ferran 1982). Contour morphostructures are concentric and radial morphostructures of a central type and are the result of structural-coordinate net analysis (Lastochkin 1991).

The criteria for detachment, contouring and determination of morphotectonic regions and areal morphostructures are their exposure in orography, spatial relations with linear morphostructures that confine them, and relationship to contours from structural-morphometric maps (Lastochkin 2006): maps of polyapexial and polybasis surfaces of the Antarctic; maps of the differences between polyapexial and polybasis surfaces; maps of the differences between polyapexial and subglacial-submarine surface; and maps of the differences between subglacial-submarine and polybasis surfaces.

Morphotectonic formations can be characterized by 2 components: a) geological notion of mineralogical composition and b) geomorphologic data on their external form and borders as well as the data on trend and intensity of shift in the neotectonic period. For lack of data on the Antarctic, geological notions of mapped morphostructures' mineralogy include only the information about which tectonic region (that underwent one or another tendency of its evolution during the neotectonic period) they belong to.

A combination of tectonic (Grikurov et al. 2003), structural – minerogenetic, geomorphologic zoning and orographic mapping of the Antarctic as well as notions on morphotectonics (Ufimtsev 2002) and Mesozoic tectonics (Bogolepov 1967) allowed for the definition of morphotectonic regions (biggest forms of subglacial – submarine surface of the Antarctic caused by neotectonic movements): areas of continental row (epiplatform orogens, piedmont plains, platform plains, subsidence plains on the block-folded basement), of geosynclinal-folded row (epigeosynclinal orogens and volcanic belt, epigeosynclinal shelves, slopes of transitional zones from continent to ocean), passive type continental margins (shelves, continental bench and continental slope), and oceanic rows (oceanic basins, middle-oceanic and block-volcanic ridges). Morphotectonic regions are presented by well-distinguished categories and by transitional units that occupy an intermediate position between first ones. They are, for example, piedmont plains separating orogens from platform plains on a continent. Continental benches include transitional areas between continental regions and continental slope. Shelf regions are characterized into 2 different types of shelves: 1) shelves of continental platform and Mesozoic-Cenozoic sedimentary basins and 2) shelves of a geosynclinal-folded type.

A continental row consists of the ancient east Antarctic craton regions; epiplatform orogens and subsidence plains on the block-folded basement of the Transantarctic province, which are transitional to the Pacific geosynclinal-folded province.

CONCLUSIONS

This research presents the first detailed description of the morphotectonic plan of the Antarctic (first scheme of neotectonics dates back in 1970s). Investigation results provide the basis for future studies on tectonics, lithodynamics and glaciodynamics.

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The Peribaltic Group is an INQUA (International Union for Quaternary Research) research group, bringing together Quaternary researchers from countries around the Baltic Sea. The 2006 meeting of the working group took place from 11th to 15th September in Finland, at the Oulanka research station in Kuusamo. Researchers came together to discuss and share the latest research results about climate changes during the Weichselian Ice Age, the movement of glaciers, the morphological features they formed and the development of the vegetation. Totally 48 lectures were given. This publication contains the peer reviewed papers of 21 lectures.

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