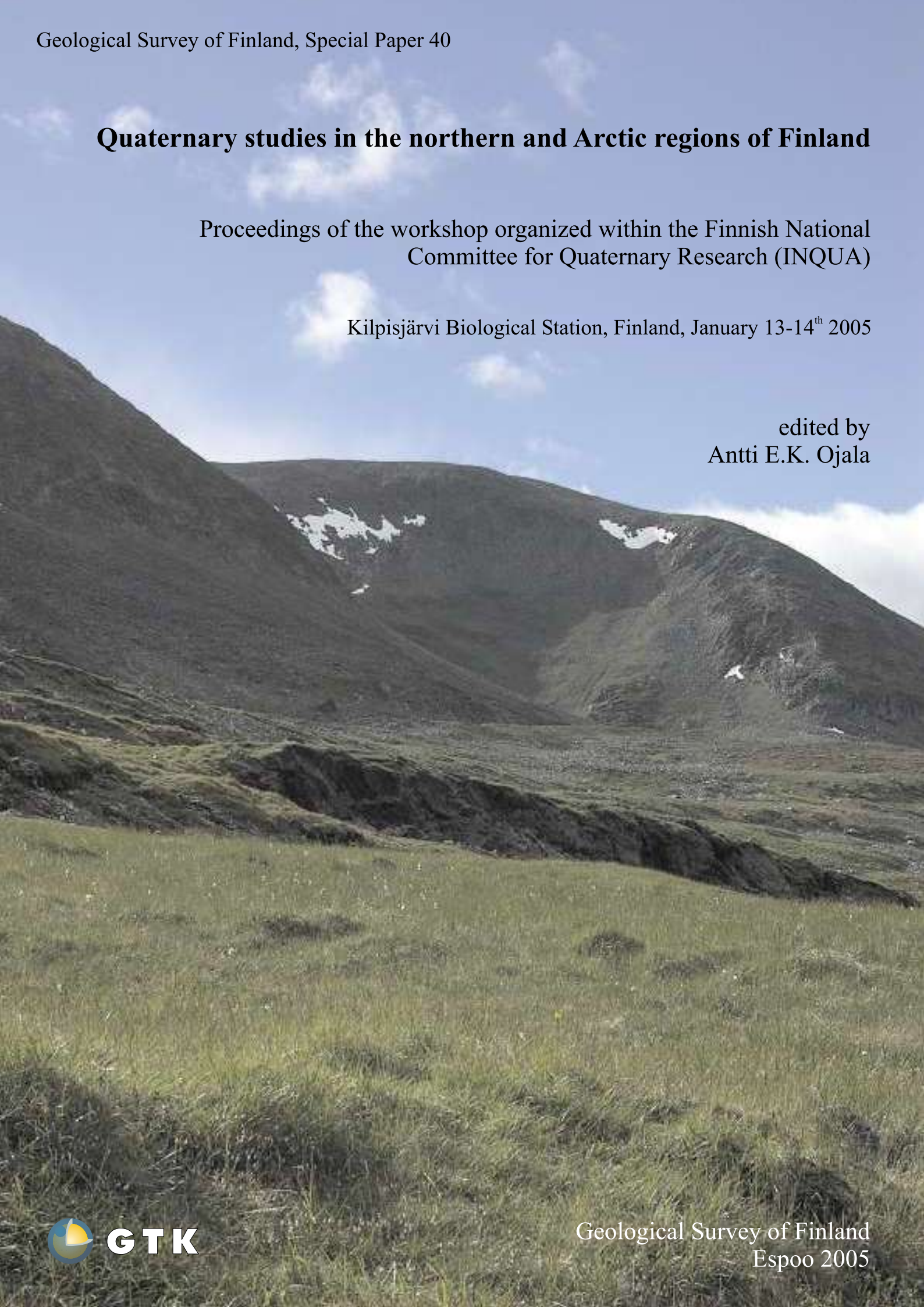
Geological Survey of Finland, Special Paper 40

Quaternary studies in the northern and Arctic regions of Finland

Proceedings of the workshop organized within the Finnish National Committee for Quaternary Research (INQUA)

Kilpisjärvi Biological Station, Finland, January 13-14th 2005

edited by
Antti E.K. Ojala



GTK

Geological Survey of Finland
Espoo 2005

Front cover: Cirque on the eastern flank of Govddosgaisi Fell (Photo: A.E.K. Ojala 2005)

Back cover: End moraine in the Govdajohka Valley 4 km suoth from Halti Fell (photo: A.E.K. Ojala 2005)

Geological Survey of Finland, Special Paper 40

Quaternary studies in the northern and Arctic regions of Finland

Proceedings of the workshop organized within the Finnish National Committee for Quaternary Research (INQUA)

Kilpisjärvi Biological Station, Finland, January 13–14th 2005

edited by Antti E.K. Ojala

Geological Survey of Finland
Espoo 2005

Ojala, A.E.K. (ed.) 2005. Quaternary studies in the northern and Arctic regions of Finland Geological Survey of Finland, Special Paper 40, 130 pages, 98 figures, 10 tables.

This publication contains 14 peer-reviewed papers that are based on presentations given at the workshop "Quaternary studies in the northern and Arctic regions of Finland" organized by the Finnish National Committee for Quaternary Research (INQUA) at Kilpisjärvi Biological Station, Finland, on 13-14 January 2005. The papers represent an overview of recent and actively ongoing Quaternary research in the Arctic and Subarctic regions of Finland.

The first paper by Hirvas et al. introduces some of the unique geological characteristics of NW Finnish Lapland. Vanhala et al. and Ojala et al. then discuss two of these geological features in more detail: the occurrence and nature of the permafrost around Ridnitšohkka Fell and the activity of Halti Fell ice tongue, respectively. Likewise, papers by Johansson, Sutinen and Koskinen are based on late Glacial/early Holocene geological investigations of local geological sites of particular interest, whereas Sarala and Mäkinen discuss the larger-scale development and dating of Weichselian Quaternary deposits in Finnish Lapland.

Seppälä outlines the main problems in the dating of palsas, one type of peatland in Finnish Lapland that is characterized by high mounds with a permanently frozen core. Hagberg presents an ongoing high-resolution study of the pollen-climate and pollen-vegetation relationships at Nellim in northern Lapland.

Moving a little southwards, a paper by Pitkäranta suggests well-considered lithostratigraphical names for different units of the Suupohja region in southern Ostrobothnia, and Breilin et al. present the unique geological characteristics of the Kvarken Acrhipelago in the western coast of Finland.

Rankama & Kankaanpää bring an archaeological viewpoint into this publication from a survey conducted around Lake Vetsijärvi (Utsjoki) to investigate the history and prehistory of the area. Finally, Lehtinen & Luttinen present results from a geochemical investigation of Lake Basen, a frozen lake located in western Dronning Maud Land, Antarctica.

Key words (GeoRef Thesaurus, AGI): glacial geology, glacial features, periglacial features, Arctic region, Quaternary, symposia, Finland

Antti E.K. Ojala
Geological Survey of Finland
P.O. Box 96, FIN-02151, Espoo, FINLAND

E-mail: antti.ojala@gtk.fi

ISBN 951-690-930-2
ISSN 0782-8535

Vammalan Kirjapaino Oy 2005

Quaternary studies in the northern and Arctic regions of Finland
Edited by Antti E.K. Ojala
Geological Survey of Finland, Special Paper 40, 2005.

CONTENTS

PREFACE

Northern and Arctic regions are of special importance in the Earth's system because they are among the last pristine environments, and are known to be highly sensitive to environmental and climate change. These areas also maintain a variety of natural archives that are suitable for basic and applied quaternary research. Evidence of longer-term natural climate variability, i.e. over the past interglacial-glacial cycles, exists in ice core records (e.g. Petit et al. 1999) and in Quaternary deposits (e.g. Svendsen et al. 2004). The urge to deepen the understanding of the Holocene climate derives from the most recent global warming, and therefore a number of studies have been targeted at investigating decadal to millennial climate variability in northern and Arctic regions. Such studies are using instrumental records and monitoring installations to examine the recent past, and natural climate archives, such as tree rings and lake and mire deposits, to go further back in time. However, global change is more than climate change. For example, long-distance transportation of contaminants into the Arctic as a major environmental threat, and the relationship between humans and the changing Earth's environment are key issues under international investigation at the moment (e.g. ACIA 2004).

Environment and climate change issues have been typical subjects of Finnish Arctic and Subarctic research over the past few decades. A geochemical baseline study on "Barents Ecogeochemistry" led by the Geological Survey of Finland (Salminen 2005) is one example of such work, together with active climate research examining both long- and short-term fluctuations (e.g. Helama et al. 2002, Seppä & Birks 2002, Lunkka et al. 2004).

This publication is based the workshop "Quaternary studies in the northern and Arctic regions of Finland" organized by the Finnish National Committee for Quaternary Research (INQUA) at Kilpisjärvi Biological Station, Finland, on 13–14 January 2005. This national workshop sought to bring together an interdisciplinary group of people working on different scientific issues in the Finnish Arctic and Subarctic region, as well as other similar regions. It aimed to engage biologists, geologists, geographers and archaeologists in studying global change and the relationship between humans and the changing Earth's system. Papers in this publication represent a selection of the scientific outcomes of the workshop in the form of presentations and discussions (see Ojala 2005 for the full content of the presentations). Papers were reviewed by Professors Matti Saarnisto and Matti Eronen, and their contribution is gratefully acknowledged. Roy Siddall kindly revised the language of many of the articles in this publication. Finally, the Finnish National Committee for Quaternary Research expresses its special gratitude to the Geological Survey of Finland (GTK) for supporting the publication of the workshop papers in this GTK Special Paper, volume 40.

Antti E.K. Ojala

Espoo, 2.9.2005

REFERENCES

ACIA 2004. Impacts of warming Arctic: Arctic Climate Impact Assessment. Cambridge: Cambridge University Press, 140 p.

Helama, S., Lindholm, M., Timonen, M., Meriläinen, J. & Eronen, M. 2002. The supra-long Scots pine tree-ring record for Finnish Lapland: Part 2, interannual to centennial variability in summer temperatures for 7500 years. In: Briffa, K.R. & Matthews, J.A. (eds.), Analysis of dendrochronological variability and associated natural climates in Eurasia (ADVANCE-10K). The Holocene 12, 681-687.

Lunkka, J.P., Johansson, P., Saarnisto, M. & Sallasmaa, O. 2004. Glaciation of Finland. In: Ehlers, J. & Gibbard, P.L. (eds.), Quaternary glaciations: extent and chronology. Part 1: Europe. Developments in Quaternary Science 2, 93-100.

Ojala, A.E.K. (ed.) 2005. Pohjoisten alueiden kvartääritutkimuksen seminaari, Abstraktikokoelma. Kilpisjärven biologinen asema, 13-15.1.2005, Geological Survey of Finland, Report P 22.4.023, 68 pp. [in Finnish]

Petit, J.R., Jouzel, J., Raynaud, D., Barkov, N.I., Barnola, J.M., Basile, I., Bender, M., Chappellaz, J., Davis, J., Delaygue, G., Delmotte, M., Kotyakov, V.M., Legrand, M., Lipenkov, V.Y., Lorius, C., Pepin, L., Ritz, C., Saltzman, E. & Stievenard, M. 1999. Climate and Atmospheric History of the Past 420,000 years from the Vostok Ice Core, Antarctica. Nature 399, 429-436.

Salminen, R. (ed.) 2005. GEOCHEMICAL Atlas of Europe BD: Part 1: Background information, methodology and maps. Geological Survey of Finland, 526 p.

Seppä, H. & Birks, H.J.B. 2002. Holocene climate reconstructions from the Fennoscandian tree-line area based on pollen data from Toskaljavri. Quaternary Research 57, 191-199.

Svendsen, J.I., Alexanderson, H.I., Astakhov, V.I., Demidov, I., Dowdeswell, J.A., Funder, S., Gataullin, V., Henriksen, M., Hjort, C., Houmark-Nielsen, M., Hubberten, H.W., Ingólfsson, Ó., Jakobsson, M., Kjær, K.H., Larsen, E., Lokrantz, H., Lunkka, J.P. Lyså, A., Mangerud, J., Matiouchkov, A., Murray, A., Möller, P., Niessen, F., Nikolskaya, O., Polyak, L., Saarnisto, M., Siegert, C., Siegert, M.J., Spielhagen, R.F. & Stein, R. 2004. Late Quaternary ice sheet history of northern Eurasia. Quaternary Science Reviews 23, 1229–1271.

Quaternary studies in the northern and Arctic regions of Finland
Edited by Antti E.K. Ojala
Geological Survey of Finland, Special Paper 40, 7–12, 2005.

GEOLOGICAL CHARACTERISTICS OF THE HALTI-RIDNITŠOHKKA REGION, ENONTEKIÖ, FINLAND

by
Heikki Hirvas, Petri Lintinen, Antti E.K. Ojala & Heikki Vanhala

Geological Survey of Finland, P.O. Box 96, FIN-02151, Espoo, FINLAND
E-mail: heikki.hirvas@gtk.fi, petri.lintinen@gtk.fi, antti.ojala@gtk.fi, heikki.vanhala@gtk.fi

Hirvas, Heikki, Lintinen, Petri, Ojala, Antti E.K. & Vanhala, Heikki, 2005. Geological characteristics of the Halti-Ridnitšohkka region, Enontekiö, Finland. *Geological Survey of Finland, Special Paper 40*, 7–12, 6 figures.

At present, the harshest climatic conditions in Finland are found at the highest fell summits of the Halti-Ridnitšohkka area, north-west of Enontekiö municipality in Lapland. The summit region of Ridnitšohkka is an extensive area representing morphological features related to recent periglacial activity. The basal till at the summit plateau is devoid of vegetation cover and displays patterned ground features. At the end of July during several summer seasons, frozen ground conditions have been found in the till at a depth of 0.85–1.9 m. Typical morphological features of the slopes are solifluction terraces and lobes. The eastern flank of Ridnitšohkka has the most extensive area of permanent snow (3 km^2) in Finland.

The present thickness of the snowfield is about 6 m, and according to radiocarbon dating of plant remnants found in the snow at a depth of about 4 m, the age of the snow is only a few decades. According historical records and aerial photographs, the snowfield has existed for at least one hundred years, but during the last few years the size has considerably diminished. During deglaciation of the Weichselian ice sheet the ice terminus retreated to the south, leaving small mountain glaciers in the area. The most distinct morphological feature is the cirque on the eastern flank of Kovddoskaisi. The valley glacier that occupied the Kovdajohka valley is well indicated by a set of more than ten end moraines. Surface boulders and stones of olivine gabro type, originating from the Halti fell, also indicate the extent and ice-flow direction from north to south, i.e. opposite to the main ice flow direction of the Scandinavian ice sheet in the area.

Key words (GeoRef Thesaurus, AGI): periglacial features, patterned ground, permafrost, till, solifluction, end moraines, cirques, Enontekiö, Finland

Introduction

The Ridnitšohkka fell is located in the north-western extremity of the Enontekiö district, in Finnish Lapland, adjacent to the Norwegian border (Fig. 1). The altitudes of the summits of Halti and Ridnitšohkka fells are 1 328 m a.s.l. and 1 317 m a.s.l., respectively, and they are the highest summits in Finland. This region differs in many respects from other parts of Lapland. The bedrock primarily consists of allochthonous sheets of early Palaeozoic rocks that were thrust south-eastwards over the Fennoscandian shield and a thin Cambrian sedimentary sequence during the Caledonian orogeny, some 400–500 million years ago. The allochthonous sheets dip gently towards the northwest and extend for only about 10–20 km

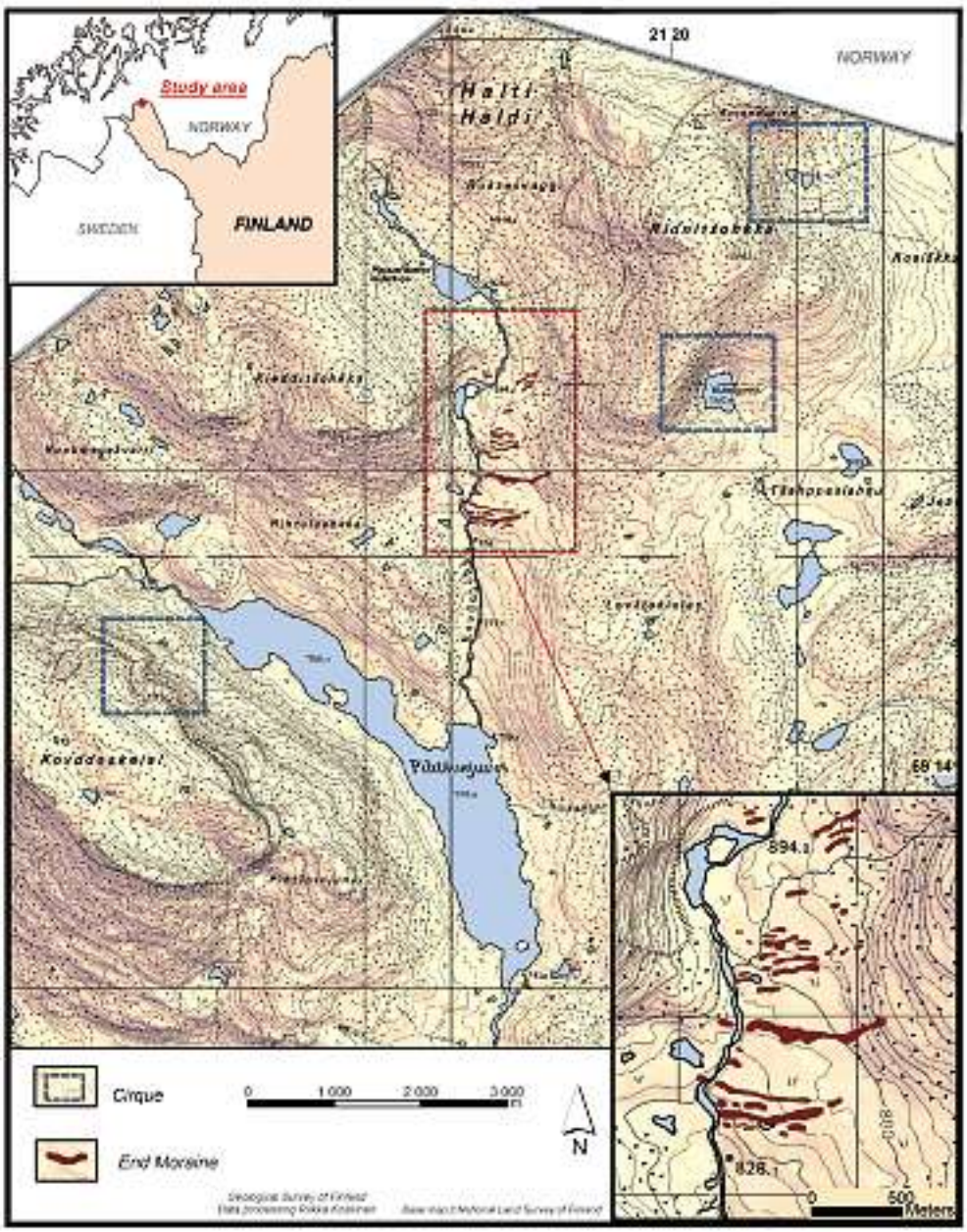


Fig. 1. Location of the study area in Lapland, Northern Finland. Three distinct cirque forms and ten end-moraines are marked in the figure, providing morphological evidence that cirque glaciers and valley glaciers existed after the deglaciation of the Weichselian ice sheet in the Kovddoskaisi-Halti-Ridnitšohkka area.

into Finland from the Norwegian border (cf. Simonen 1980, Lehtovaara 1995). The Kilpisjärvi-Halti area contains all of the Finnish peaks that have elevations greater than 1000 m above sea level. Relatively broad, steep-sided valleys separate the higher peaks from one another, resulting in the greatest topographical relief to be found in Finland, with 200–500 m differences in elevation between summits and valley floors being typical. The most extreme example is on the north-eastern flank of Kovddoskaisi, with a steep rock face more than 500 m high. Various morphological features in the Halti-Ridnitšohkka area are indicative of extensive periglacial activity.

After the deglaciation of the Scandinavian ice sheet, one valley glacier and three cirque glaciers existed in the area. At present the area is characterised by the harshest climatic conditions in Finland, as indicated by the largest permanent snowfield, 3 km^2 in size, and an extensive area representing morphological features related to periglacial activity. Climatologically the region belongs to the area of discontinuous permafrost delineated with the presence of palsa mires (King & Seppälä 1987).

Halti-Kovdajohka valley glacier

During the deglaciation of the Weichselian ice sheet in the Halti region, the terminus retreated roughly

Fig. 2. End moraine in Kovdajohka valley. The ice flow direction is from left to right. (Photo: P. Lintinen, 1991)

from north to south. After the deglaciation of the ice sheet a valley glacier occupied the Kovdajohka valley (Kujansuu 1967), which so far has been the only valley glacier known to be active in the present territory of Finnish Lapland (Tanner 1915). The valley glacier flowed from Halti towards south via the Kovdajohka valley as a six kilometre long tongue, producing more than ten end moraines at the altitudes of 830–930 m a.s.l. (see Fig. 1). The end moraines vary in size and morphology, the southern most ones being distinctively greater in size and representing a steeper-sided ridge-like morphology than the northern ones (Fig. 2). The most prominent end moraine extends across the whole valley, being about 900 m in length, 4–5 m in height and 50–60 m in width. In both till material and surface boulders, the end moraines and their adjacent areas are rich in olivine gabro originating from Halti (Kujansuu 1967, Hirvas 1968). Boulders of olivine gabro type do not exist outside of the end moraines. Olivine gabro is found in bedrock only in the Halti fell area, which evidently shows that the valley glacier has flowed from north to south. This direction is opposite to the main south to north ice flow direction of the ice sheet in the area before deglaciation.

Three distinct cirque forms provide morphological evidence that cirque glaciers existed after the degla-

Fig. 3. The most prominent cirque in Finland, located in the north-eastern wall of Kovddoskaisi, adjacent to Lake Pihtsusjärvi, about 6 km south-west from Ridnitšohkka. The location is marked in Fig. 1. The back wall of the cirque is about 300 m in height and 500 m in width. Perennial snow is found in the cirque. The photograph was taken at the end of July 2004. (Photo: H. Vanhala, 2004)

Fig. 4. An oblique aerial photograph from the eastern slope of Ridnitšohkka at the end of July 1991. (Photo: P. Lintinen, 1991)

ciation of the ice sheet in the area. Two of the cirque glaciers existed on the eastern flank of Ridnitšohkka fell and the third on the north-east slope of Kovddoskaisi fell (Fig. 3).

Ridnitšohkka snowfield

A continuous snowfield is present at elevations of between 1270–1180 m a.s.l. on the northern and eastern flanks of Ridnitšohkka fell. The earliest written records documenting the presence of a large permanent snowfield near the summit of Ridnitšohkka date from the latter part of 19th century (Stjernvall 1892). The snowfield has most probably existed since the earliest descriptions. Aerial photographs taken in 1960, 1969, 1975, 1979, 1985 and 1991 show that the snowfield has had approximately the same size and shape during the last 30 years. At the beginning of August 1991, one continuous snowfield had a maximum width of 600 m and a length of nearly 5 km (Fig. 4). Drilling at the deepest locality, chosen by ground penetrating radar sounding, revealed a depth of 6.2 m.

The density of the snow and ice samples drilled from the snowfield varied between 540 – 670 kg/m^3, which according to Patterson (1994) is typical for perennial snow, or firn. An icy surface rich in organic material, remnants of grass, was found at the depth of 4.05 m. This evidently represents the remnant of an earlier surface of the snowfield with wind blown plant remains, which were revealed by AMS radiocarbon dating (Ua-2886) to be of recent age. The high radiocarbon activity of the studied sample was equal to the radioactivity of the atmosphere during 1957–58 and indicated that the plants had been living when atmospheric nuclear explosion tests were conducted. Independent and somewhat surprising support for the young age of the surface was the finding of pieces of styrofoam ("styrox") together with the plant remnants. In Finland the application of styrofoam as a thermal insulating material in building construction became common after 1952. According to Hirvas et al. (2000), the turnover time of the snow and ice in the snowfield must be relatively short, only a few tens of years. During the winter, snow accumulates on top of the snowfield, whereas during the summer the base of snow and ice melts, resulting in streams that run between frost weathered bedrock and the snowfield.

Comparing the present extent of the snowfield at the end of July 2004 to observations conducted at the end of July 1991 and to the aerial photographs taken in late summer 1991, the size and shape of the snowfield have decreased considerable. In the area of the former large snowfield, a few smaller separate snowfields existed at the end of July 2004 (Figs. 4 and 5). One possible reason may be that during several recent winters the snow cover in the Kilpijärvi-Halti region has been exceptionally thin (oral communication by Finnish Border Guard, Kilpisjärvi base).

Fig. 5. An oblique aerial photograph from the eastern slope of Ridnitšohkka at the end of July 2004. The greyish area adjacent to the snowfields is the zone of bare rocks and block fields free of lichen and moss, recently exposed from the snowfield. (Photo: A.E.K. Ojala, 2004)

Patterned ground and permafrost

Patterned ground devoid of vegetation, including sorted and unsorted polygons and stone stripes, are typically found in the summit areas and on the flanks of fells in the research area. On the summit plateau of Ridnitšohkka, the gently sloping till cover exhibits well developed sorted polygones free of vegetation at the altitude of 1 290 m a.s.l. The disposition of overturned boulders and stones, and the presence of patches of disaggregated till material on top of some boulders, indicates that periglacial processes are actively operating in the area. Drilling with a portable Cobra-percussion drill equipped with a 25 mm flow-through bit was carried out at two sites on the summit plateau on 29–30 July 1993. The two sites were situated in patterned ground and spaced about twenty metres apart. The individual drilling sites were situated in the centre of polygons that were devoid of vegetation cover. In both cases the surficial part of the till cover was extremely wet and easy to drill. The penetration ceased at depths of 1.97 and 2.0 m. In both cases the drill bits contained 7–10 cm pieces of frozen till containing numerous subparallel, gently dipping laminae of 1–2 mm thick ice (Hirvas et al. 2000). At the same site, in early August 2004, frozen till was noticed at the depth of 85 cm (Vanhala et al. in press).

Solifluction terraces

A series of distinctive solifluction terraces have developed on slopes where the till cover is somewhat thicker. On the eastern flank of Ridnitšohkka, at an elevation between 970–850 m, there are some 30 nearly subparallel till terraces, of which the largest are up to 4 m high and more than a kilometre in length (Hirvas 1968, Hirvas et al. 2000). The uppermost parts of these till terraces are free of plant cover and very rocky, whereas lower down they are covered by vegetation and have a typical lobate morphology (Fig. 6). The solifluction terrace/lobes, lowest in altitude, coincide with a line of springs, which may be an indication that the deeper parts of the till sequence are frozen and therefore rain and melt waters do not infiltrate deeper into the till.

Conclusion

The Halti-Ridnitšohkka region is unique in Finnish Lapland in terms of altitude, relief, bedrock and glacial geology and periglacial processes. Subsequent to the deglaciation of the Scandinavian ice sheet, a valley glacier and three cirque glaciers were active in the area. At present, on the summit plateau of Ridnitšohkka fell, active periglacial processes operate on a layer of perennially frozen till, resulting in patterned ground devoid of vegetation. The most extensive permanent

Fig. 6. An example of solifluction lobes/terraces on the eastern slope of Ridnitšohkka. One individual terrace, situated perpendicular to the dip of the slope, can be traced in aerial photographs for more than 1 km. On a minor scale the terraces have a morphology of numerous lobes. (Photo: A.E.K. Ojala, 2004)

snowfield in the territory of Finland exists on the eastern flank of Ridnitšohkka, together with great number of solifluction terraces, hundreds of meters in length, some of which are still active.

Acknowledgements

The Lapland Frontier Guard District and Unibase Oy provided important accommodation and logistics support during the spring and summer 2004–2005 fieldwork.

REFERENCES

Hirvas, H. 1968. Haltin alueen Kvartäärisestä kehityksestä. University of Helsinki, Department of Geology. Unpublished M.Sc. thesis, 67 p.

Hirvas, H., Lintinen, P. & Kosloff, P. 2000. An extensive permanent snowfield and the possible occurrence of permafrost in till in the Ridnitšohkka area, Finnish Lapland. Bulletin of the Geological Society of Finland 72, 47–56.

King, I. & Seppälä, M. 1987. Permafrost thickness and distribution in Finnish Lapland – results of geoelectrical soundings. Polarforshung 57, 127–147.

Kujansuu, R. 1967. On the deglaciation of western Finnish Lapland. Bulletin de la Comission Géologique de Finland 232, 98 p.

Lehtovaara, J.J. 1995. Kilpisjärven ja Haltin kartta-alueiden kallioperä. Summary: Pre-Quaternary rocks of the Kilpisjärvi and Halti map-sheet areas 1:100 000, Explanation to the Maps of Pre-Quaternary Rocks, Sheet 1823 and 1842, Geological Survey of Finland, 64 p.

Patterson, W.S.B. 1994. The Physics of Glaciers. Third edition. Oxford: Elsevier, 480 p.

Simonen, A. 1980. The Precambrian in Finland. Geological Survey of Finland, Bulletin 304, 58 p.

Stjernvall, H.J. 1892. Bidrag till Finska Lappmarkens geognosi II. Försök till framstälning af de geologiska förhållandena mellan Könkämäeno och norska gränsen. Medd. Från Industristyrelsen 17, Helsinfors.

Tanner, V. 1915. Studier öfver kvartärsystemet I Fennoskandias nordliga delar III. Résumé en francais. Bulletin de la Comission Géologique de Finland 38, 815 p.

Vanhala, H., Suppala, I., Lintinen, P., Hirvas, H. & Ojala, A.E.K. Application of electrical and electromagnetic methods for studying frozen ground and bedrock – results from Ridnitšohkka, Northern Finland. Geological Survey of Finland, Special Paper, in press (this publication).

Quaternary studies in the northern and Arctic regions of Finland
Edited by Antti E.K. Ojala
Geological Survey of Finland, Special Paper 40, 13– 22, 2005.

APPLICATION OF ELECTRICAL AND ELECTROMAGNETIC METHODS IN STUDYING FROZEN GROUND AND BEDROCK – RESULTS FROM RIDNITŠOHKKA, NORTHERN FINLAND

by
Heikki Vanhala, Ilkka Suppala, Petri Lintinen, Heikki Hirvas & Antti E.K. Ojala

Geological Survey of Finland, P.O. Box 96, FIN-02151, Espoo, FINLAND
E-mail: heikki.vanhala@gtk.fi, ilkka.suppala@gtk.fi, petri.lintinen@gtk.fi, heikki.hirvas@gtk.fi, antti.ojala@gtk.fi

Vanhala, Heikki, Suppala, Ilkka, Lintinen, Petri, Hirvas, Heikki & Ojala, Antti E.K., 2005. Application of electrical and electromagnetic methods in studying frozen ground and bedrock – results from Ridnitšohkka, Northern Finland. *Geological Survey of Finland, Special Paper 40*, 13–22, 8 figures and 2 tables.

We present results from the use of resistivity soundings and airborne electromagnetic techniques for mapping frozen ground and permafrost in the Ridnitšohkka fell area in Northern Finland. Permafrost is a very typical phenomenon in Arctic and Antarctic regions. In regions where permafrost alternates with seasonal frost, the permafrost typically exists discontinuously. Especially in coastal areas and near lakes and glaciers, the thickness of the active layer and permafrost vary considerably. In Finland, the northernmost part of the country is regarded as a discontinuous permafrost area. The southern limit of the discontinuous permafrost is usually delineated by the southern limit of distribution of palsa mires. The only direct evidence of perennially frozen minerogenic soil, or ground, in Finland is from the Ridnitšohkka study area.

The ground resistivity data indicated that the till cover in the summit area of Ridnitšohkka is frozen to a depth of several metres. The thickness of the thawed till, the active layer, at the end of July 2004 was close to one metre. The electrical structure derived from airborne electromagnetic data was very similar to the structure measured by ground resistivity soundings. This finding may extend the use airborne EM techniques in regional scale permafrost studies in regions of thin or discontinuous permafrost.

Key words (GeoRef Thesaurus, AGI): permafrost, frozen ground, till, bedrock, geophysical methods, electrical sounding, resistivity, electromagnetic methods, Ridnitšohkka, Enontekiö, Finland

Introduction

Permafrost, perennially frozen ground, is typical in Arctic and Antarctic regions. Almost a quarter of the Earth's land area is coved by permafrost. In Canada and Russia, half of the land area is permafrost covered. The active layer of permafrost is its annually thawing top layer. The active layer is characterised by strong annual changes in soil properties, including oversaturated water content during summer seasons and volume changes in freezing conditions. The thickness of permafrost may be up to several hundreds of metres while the thickness of the active layer is always less than two metres. In regions where

Fig. 1. A map showing the area of palsa mires, i.e., the area of sporadic permafrost, (Seppälä 1997) and the location of the Ridnitšohkka study area in Northern Finland.

permafrost gradually alternates with seasonal frost, the permafrost typically exists discontinuously in favourable locations. Especially in coastal areas and near to lakes and glaciers, the thickness of the active layer and permafrost varies considerably even within short distances. In Arctic regions a vast number of dwellings and industrial buildings, roads and constructions related to the oil industry, mining and so on have been built on frozen ground. In such locations, even partial melting of the permafrost can cause serious problems for the infrastructure and environment due to the failure of foundations.

The presence of palsa mires is indicative of permafrost in Finnish Lapland (Fig. 1). The southern limit of discontinuous permafrost is often delineated by the southern limit of distribution of palsas (Seppälä 1997). It can be traced across northernmost Finland from just north of Lake Inari through Hietatievat and further westwards at a latitude of 68° 25' to the Swedish border (King & Seppälä 1987, Seppälä 1997). In palsa mounds the low thermal diffusivity of peat restricts the active layer thickness during summer seasons. Continuous evaporation due to the moisture content in the active layer helps the frozen core of a palsa mound to remain frozen over the summer season. The only direct evidence of perennially frozen minerogenous soil and ground in Finland has resulted from the drilling of a till layer at the summit of Ridnitšohkka fell (Hirvas et al. 2000, Hirvas et al. in press). Morphological features related to frost processes, such as distinct solifluction lobes and vegetation-free patterned ground, suggest that the frost action is intense in extensive areas in fell summits and on the highest slopes and valley floors. In Peera, where the Kilpisjärvi road runs through a palsa mire, the minerogenic ground is expected to be frozen throughout the year due to the snow-free road surface allowing the cold to penetrate to deeper layers than with snow-covered soil surfaces (Saarelainen 1990). Indirect geophysical evidence also points to the existence of permafrost in the minerogenic layer underlying palsa mounds in Peera (Vanhala et al. 2001, Vanhala & Lintinen 2002).

The objective of this study was to test how the extent and internal structure of seasonal or perennially frozen ground can be mapped by geophysical techniques, namely airborne electromagnetic (EM) and ground electrical measurements. The selected research area was the summit and western flank of Ridnitšohkka fell. The summit area was chosen because perennially frozen conditions have been shown to prevail there in the till layer for a period of several years, pointing to the existence of permafrost in the mineral soil.

The applicability of electrical and EM techniques in mapping frozen ground and permafrost is based on the increase in the electrical resistivity of the ground as it freezes. In unfrozen conditions, the electrical resistivity of soil and bedrock mainly depends on the concentration of dissolved charged ions in the pore water and on the porosity and in unsaturated materials on the degree of saturation.

Resistivity also depends on the temperature, so that in a case of pure water, a decrease in temperature from +25 °C to +1 °C roughly doubles the electrical resistivity. Below the freezing point, the increase in resistivity due to decreasing temperature is much stronger. According to Scott and Kay (1988), the resistivity of silt, sand and gravel increases by a factor of ten when the temperature decreases from zero to –5 °C. For fine-grained sediments, clay and clayey till, and for peat, the increase was only five-fold, or slightly less (Hoekstra & McNeill 1973, Scott & Kay 1988). The rock data presented by Olhoeft (1978) and Scott and Kay (1988) include biotite gneiss and basalt, the former displaying the same kind of behavior as sand and gravel, i.e., a ten-fold increase in resistivity (from about 5 000 Ω·m to 50 000 Ω·m) along with a decrease in temperature from zero to –5 °C. For basalt (Olhoeft 1978), the respective increase in resistivity was about 100–fold (from 5000 Ω·m to 500 000 Ω·m). In water-rich materials, clay and peat, the resistivity increased steadily with decreasing temperature, but in the case of clayey till, silt, sand and gravel the temperature-resistivity curve flattened around temperatures of –8 to –10 °C. The resistivity data discussed above are presented in Table 1. For comparison, typical Finnish values for sediments, peat and rocks are presented. The Finnish resistivity values for unfrozen materials are up to ten-fold higher than the values presented

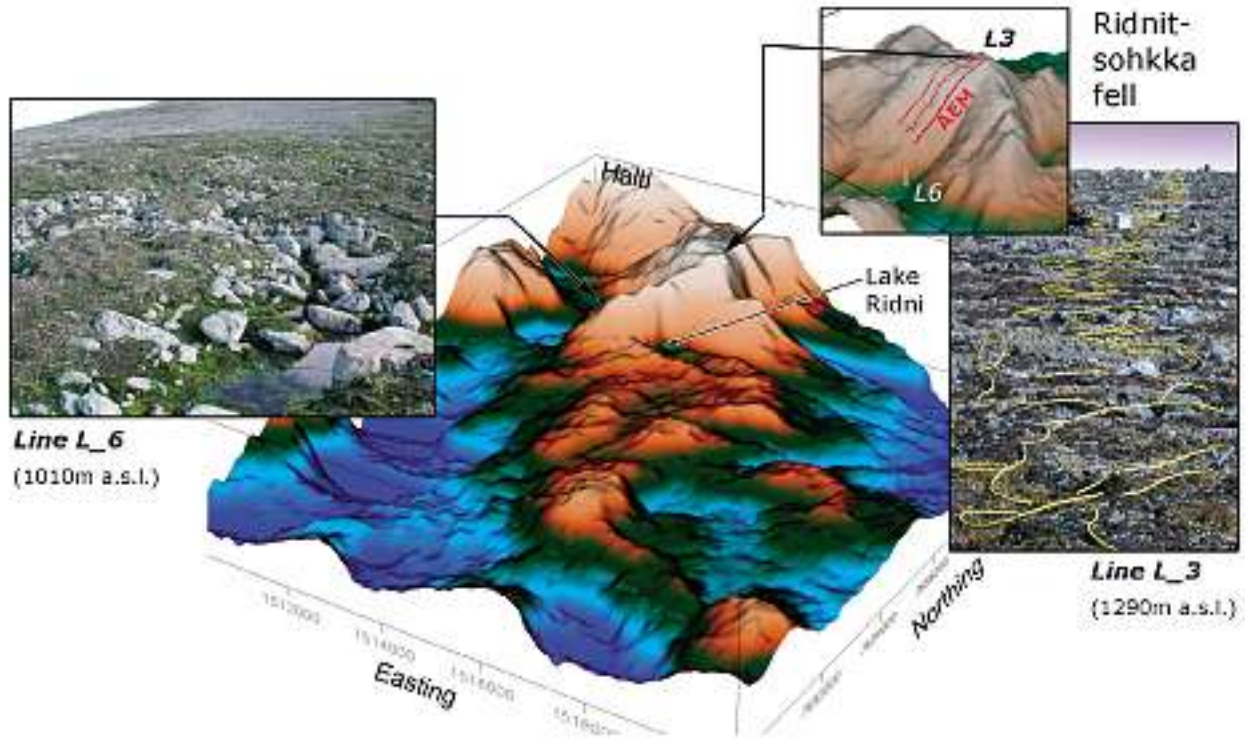


Fig. 2. Elevation model of the Halti-Ridnitšohkka area based on DGPS and radar data measured during the aerogeophysical survey (200 metres line spacing). The locations of the ground profiles (L3 and L6) and the three east-west trending airborne lines (AEM), which are discussed here in detail, are marked in the upper right figure. The map area is 10 x 10 km^2. The vertical scale is strongly exaggerated, the elevation difference between the highest summits and the valleys being about 500 metres. In the photo from line L_3, the electrode layout for resistivity soundings is illustrated.

by Hoekstra & McNeill (1973), Scott & Kay (1988) and Olhoeft (1978). No data for frozen Finnish soil and rock material are presented, but apparently the behaviour resembles that presented in Table 1.

Field resistivity soundings results from a palsa mire at Peera in northern Finland show a difference of 10- to 100-fold between unfrozen and frozen peat (Vanhala 2000, Vanhala et al. 2001, Vanhala & Lintinen 2002). In laboratory tests carried out at the Geological Survey of Finland, a nearly 100-fold difference was measured

Table 1. Resistivity values for common soil and rock types: (1) Hoekstra and McNeill (1973), Scott & Kay (1988), Olhoeft (1978), (2) Peltoniemi (1988), ***no temperature data, apparently room temperature (+20 °C).

ref.	soil/rock type	resistivity (Ω·m), +10°C/***	resistivity (Ω·m), –5°C
(1)	gravel	400	4 000
(2)		1 000–2 000	
(1)	sand	80	1 500
(2)		500–1 000	
(1)	silt	30	500
(2)		80–200	
(1)	clay	20–30	50
(2)		30–60	
(1)	clayey till	30	100
(2)	till	200–500	
(1)	peat	100	500
(2)		80–140	
(1)	basalt	2 000	50 000
(1)	gneiss	5 000	500 000
(2)	gabbro, granite, gneiss, mica schist, quartzite	10 000–20 000	

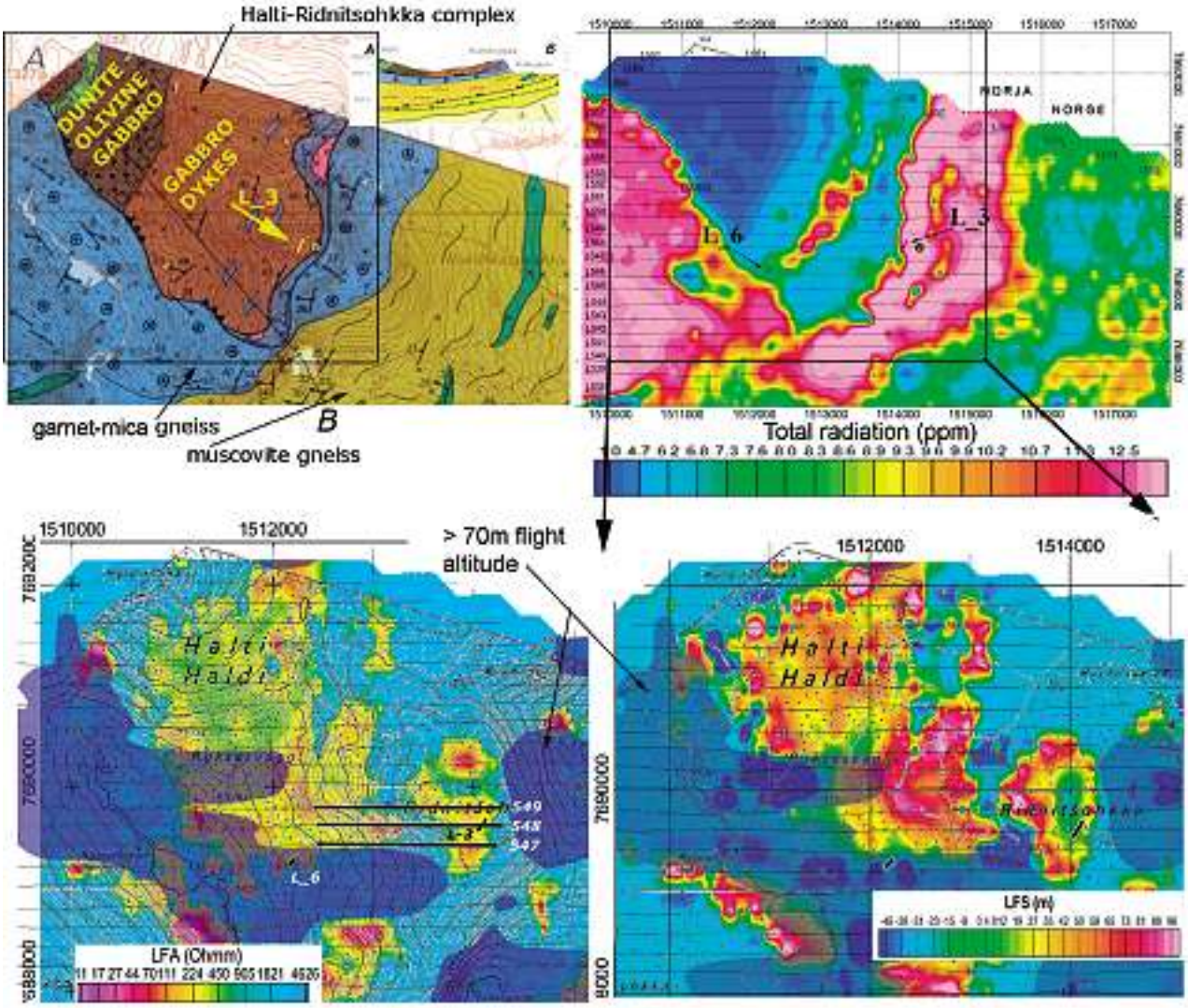


Fig. 3. A geological map (Lehtovaara 1995) in the upper left; an airborne radiometric map in the upper right (total radiation in units of ppm); an airborne conductivity map (3.1 kHz apparent resistivity, LFA) in the lower left; and an airborne apparent depth map (3.1 kHz, LFS) in lower right. The mapping flight date was 20 July 2002, the line spacing 200 m and the nominal flight altitude 30 m. Ground profiles L_3 and L_6 and flight profiles 547, 548 and 549 are marked. In the lower maps, areas with a flight altitude higher than 70 m are marked by blue transparent shading. The map area (in the lower maps) is 5x5 km^2.

for sea ice samples (700–2000 Ω·m) and water samples melted from the ice (20 Ω·m). No temperature data are, however, available from either case.

In rocks containing metallic minerals such as sulphides, magnetite or graphite the dominating electrical conductivity mechanism changes from electrolytic and the effect of freezing on the electrical resistivity disappears.

So, the electrical image to be excepted due, for example, to the ground having a frozen top layer is a section with a high-resistivity top layer. The fact that the groundwater table, the thickness of the sediments and the electrical resistivity of the bedrock are not usually known makes the interpretation generally complicated. In the case of saturated sediments, the range of resistivity is typically fairly narrow and the effect of freezing on the resistivity can be expected to be distinct. In the case of bedrock, however, the range of resistivity is wide, reaching from tens of thousands of Ohm-metres in intact rocks to tens of Ohm-metres in highly fractures rocks, and even lower in rocks containing metallic minerals.

Study area

The Ridnitšohkka fell is located just north of 69°N in the north-western extremity of the Enontekiö municipality (Fig. 1), some 40 km north-west from Kilpisjärvi. It is adjacent to Halti, the highest fell in Finland situated on the border between Finland and

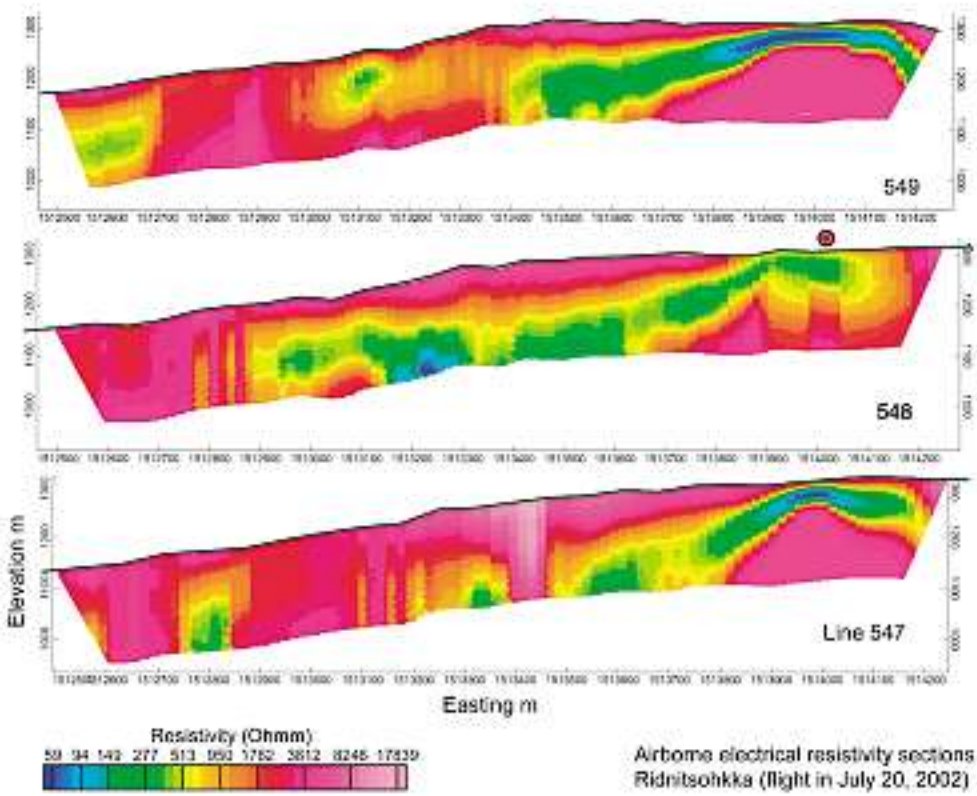


Fig. 4. Resistivity sections (see Figures 2 and 3) based on the dual-frequency airborne electromagnetic (AEM) data, 1D smooth inversion. Flight line 548 crosses the ground resistivity line L_3 (indicated by a red circle). Note the partly discontinuous low-resistivity layer at a depth of a few metres at the top of the fell, but in deeper levels on the lower slopes.

Norway. The altitude of the Ridnitšohkka fell is 1317 m a.s.l. The study areas are located at altitudes of 1290 m a.s.l and 1000 m a.s.l.

The bedrock of the Kilpisjärvi-Halti area consists of Archean basement and sheets of Caledonian orogeny trusted onto the basement and autochthonous late Proterozoic–Cambrian sediments lying on the basement. The lowermost part of the Caledonian, the Dival Group, starts with a thin layer of conglomerates on the basement. The conglomerates are covered by a sequence of sand and clay stones up to 150 metres thick. The Dival Group is overlain by parautochthonous Jerta Nappe dolomites, shales and quartzites (Lehtovaara 1995). The Dival and Jerta Group rocks are overlain by an allochthonous far-transported Kalak nappe complex, which in Finland can be divided into thee flat-laying overthrusted nappe units. From bottom to top, they are Nalganas, Nabar and Vaddas Nappes (Lehtovaara 1995). The Nalganas Nappe consists mostly of arkose quartzites, and the Nabar Nappe of serisite quartzites, mica gneisses, amphibolites and pegmatites. The top of it is monotonous garnet mica gneiss. The third and uppermost Nappe consists of a succession of ultramafic - mafic rocks - dunite, troctolite and olivine gabbro - and a sheeted dolerite complex (Andréasson et al. 2003) (a dike gabbro after Lehtovaara 1995). This unit belongs to Vaddas Nappe and together with sillimanite gneiss, palingenic granite and zoisite amphibolite forms the Halti-Ridnitšohkka complex.

The study sites (Fig. 2) are situated on the gabbroic part of the Halti-Ridnitšohkka complex (Fig. 3). The Ridnitšohkka gabbro sill unit is composed of gabbro sills of varying thickness (from tens of centimetres to tens of metres) and interlayers of garnet-sillimanite gneisses (Sipilä 1992, Lehtovaara 1987). The gneisses occur as fragments or concordant layers between the sills. Their thickness is similar to sills, from centimetres to tens of metres. The number of gneiss interlayers is low near to the peridotite–olivinegabbro cumulates, but increases by up to 30–40 percent further from the contact. In the radiometric map (Fig. 3) the Halti-Ridnitšohkka complex is characterized by a low to very low radiation level, while the garnet-mica-gneiss of the Nabar Nappe sheet delineates as a distinctively highly radiating unit. The higher-radiating anomaly inside the gabbro sill unit may arise from a high proportion of gneiss inclusions.

The summit of Ridnitšohkka is relatively flat and dips gently to the northwest. Most of the summit area is composed of fragmented angular rocks, which

occasionally alternate with more unbroken bedrock outcrops. Close to the summit also exists a till covered area about two square km in size. The till surface is composed of unsorted polygons, whose diameter varies from 0.5 to 1.0 m. These polygons are free of vegetation and are therefore evidently caused by active frost and thaw action and are evidence of present periglacial processes. On the steeper slopes the morphology varies from elongated polygons to stone stripes, all showing a preferred downward direction. In the area of relatively symmetrical polygons the soil is composed of till, which, according to earlier observations, is at least 2 m in thickness (Hirvas et al. 2000).

The second study site is situated at lower altitudes on the western slope of the Ridnitšohkka, at an altitude of 1000–1010 m a.s.l. At this study site the ground surface is composed of till and dips about five degrees to the west. In the northeast part (upper slope) of the resistivity profile, the morphology of the till surface comprises polygons of 0.5–0.9 m in size. The centre parts of the polygons are covered with moss, and they are about 0.2–0.3 m above the intersections of adjacent polygons.

Methods

Two till areas were selected for representative study, with an altitude difference of about 300 m. According to the polygonal surface morphology the activity of frost and thaw action was expected to vary considerably. In middle part of each survey profile a test pit was dug by spade in order to record the physical and chemical characteristics of the till. Temperature profile was measured with a thermometer in the walls and bottom of the study pit. Chemical characteristics of soil/till samples were measured by ICP-MS after 0.01M $BaCl_2$ extraction (solid liquid ratio 1:10). The pH was also measured after 0.01M $BaCl_2$ extraction. Cation exchange capacities were calculated as a sum of base cations (Na^+, K^+, Ca^{2+}, Mg^{2+}) and acid cations (Al^{3+}, H^+). Grain-size distributions were determined with a combination of wet-sieving and sedigraf analysis. These analyses were performed in order to evaluate the correlation between soil/till characteristics and electrical properties.

Geophysical data included airborne mapping carried out on 20 July 2002 (electrical resistivity, magnetic and radiometric data), ground resistivity soundings performed in July 2004 (July 22–26) and laboratory resistivity measurements of till samples taken from the sounding sites together with the resistivity soundings. The laboratory samples were duplicates of the samples taken for chemical analysis. The resistivity soundings were performed using the AGI/Sting multi-electrode system (Wenner array, 1m, 2m or 3m electrode spacing and 24, 42 or 56 electrode groundings) and the data inverted to 2D resistivity sections (Loke & Barker 1996).

The airborne data were acquired by the dual-frequency AEM system of GTK using a Cessna Caravan aircraft. The nominal line spacing was 200 m and the nominal flight altitude 30 m. The airborne EM system (AEM-99) uses dual-frequency vertical coplanar coils. The two frequencies are 3005 kHz and 14368 kHz. The coil pairs are mounted on the wingtips so that the coil spacing is 16.96 metres. The measured quantity is the ratio H_h^{sec}/H_h^{pri}, expressed as parts-per-million (ppm), so that H_h^{sec} and H_h^{pri} are the horizontal components of the secondary and primary magnetic field. Measurements include in-phase and quadrature components (in-phase and 90° out-of-phase with respect to the primary magnetic field). A detailed described of the measuring system is given by Poikonen et al. (1998)

The in-phase and quadrature maps are the first step in the visual interpretation of the AEM data. They give direct information on the characteristics of the conductivity distribution of the ground. Transforming the EM data into the apparent resistivity (ρ_a) of a half-space (Peltoniemi 1982) is the simplest modelling technique. It assumes homogeneous ground and gives the resistivity (the apparent resistivity, ρ_a) of this ground and depth to the top of this ground (apparent depth). Apparent resistivity and apparent depth maps are calculated separately for the high and low frequencies. Especially in areas characterized by low resistivity, visual interpretation of the apparent resistivity and apparent depth maps is fruitful. In this study the second step of AEM interpretation was 1D layered-earth inversion (a smooth model norm-based inversion). In the 1D layered-earth model the earth is composed of layers each having a uniform resistivity. The 1D conductivity structure, i.e., the resistivity of each layer, is sought by regularized inversion. The goal is to find for every data point the minimum-structure model that best fits the measured data.

Results

The airborne EM data from the Halti-Ridnitšohkka area are characterized by a low-resistivity feature. The resistivity values in the low-frequency airborne (LFA) apparent resistivity map (Fig. 3) are mostly between 100–400 Ω·m and the low-resistivity anomaly roughly covers the mafic-ultramafic Halti-Ridnitšohkka complex. The corresponding apparent depth values, indicating the depth to the low-resistivity layer, are typically from a few tens of metres to 90 metres.

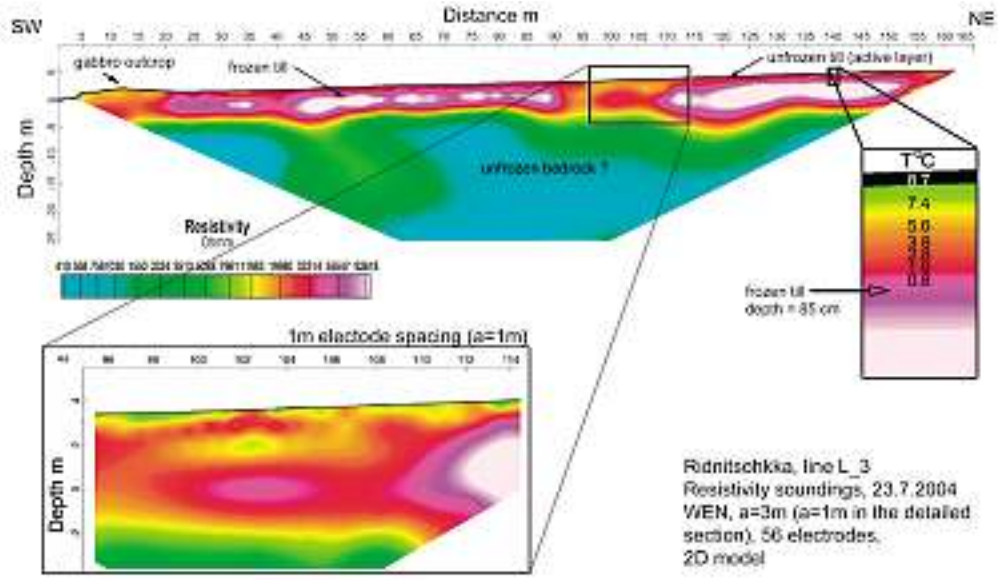


Fig. 5. Resistivity section (2D model) based on ground resistivity soundings (Wenner, a = 3m, 56 electrode groundings), Line L_3, top of Ridnitšohkka fell, 1290 m a.s.l., 23 July 2004. The detailed section from 96-114 m has been measured using one-metre electrode spacing (a = 1m). In the detailed section at right, temperatures measured in the pit are presented.

At the top of the fell, the 1D inversion of the AEM data (Fig. 4) gives a model consisting of a 10 metres thick high-resistivity (thousands of Ohm-meters) top layer and a 30 – 40 metres thick low-resistivity (~100 Ω·m) layer beneath it. The three resistivity sections in Figure 4 suggest a discontinuous low-resistivity layer that is close to the surface in the summit area, but deeper on the lower slopes of the fell. The 1D inversion results coincide with the apparent resistivity and apparent depth maps from the same site and it seems that this structure (a low-resistivity layer at a depth of tens to a hundred metres) is typical for the whole Halti-Ridnitšohkka complex area.

The ground resistivity soundings give a much more detailed image of the subsurface resisitivity distribution than the airborne EM. In the summit area of the fell, a typical electrical structure derived from resistivity soundings was (from top to bottom):

- 0.5–1 metres thick, ~10 kΩ·m, top layer
- 3–6 metres thick, 30 – 100 kΩ·m, interlayer
- 0.5–5 kΩ·m bottom layer

The 0.5–1 metres thick top layer describes thawed till (active layer). The water table was close to surface, typically at a depth of 20–30 cm, in the summit area, and the resistivity values of around 10 kΩ·m for unfrozen saturated till are extraordinarily high. They are, however, very similar to the laboratory resistivity values presented in Table 2. The slightly lower laboratory values may arise from the lack of boulders in the laboratory samples.

Comparison of the airborne data and ground resis-

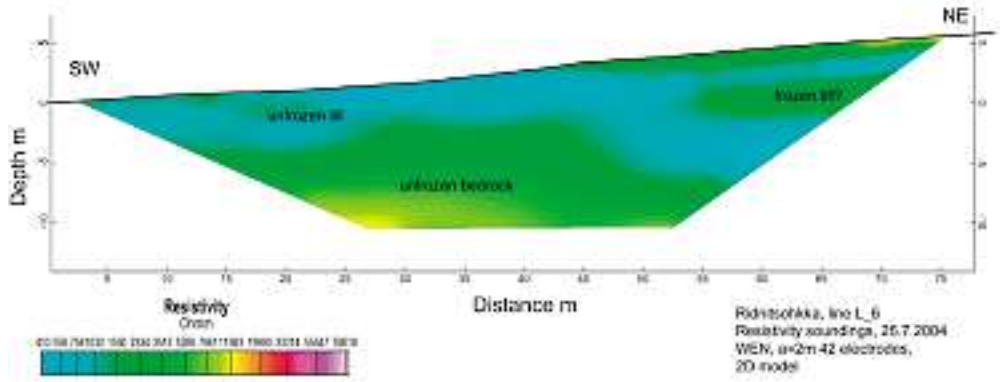


Fig. 6. Resistivity section (2D model) based on ground resistivity soundings (Wenner, a = 2m, 42 electrode groundings). Line L_6, western slope of Ridnitšohkka fell, 1000-1010 m a.s.l. (NE end), 26 July 2004.

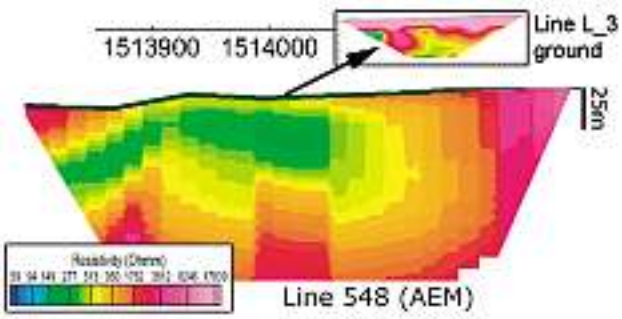


Fig. 7. Comparison of airborne and ground results, resistivity sections, Line L_3 (ground) and lie 548 (airborne EM) (see Figure 3 – lower left, and Figure 4).

tivity results is problematic because of the regional nature of the former and the very detailed nature of the latter. However, at the site where the flight line and ground line cross each other (see Fig. 3), the results are very similar (Fig. 7). The structure is similar (high-resistivity top layer and low-resistivity layer beneath it) and the resistivity values comparable. The resistivity of the top layer is markedly higher in the ground section, but this may result from the different line directions, or from the different time of measurement (AEM in 2002, ground in 2004). Additional reasons are the smooth AEM inversion and the differences between the ground and AEM techniques. The AEM-99 system is at its best between 1 and 100 Ω·m, but completely insensitive when the resistivity is over 10 000 Ω·m.

Observations made in a pit dug on the resistivity line L-3 revealed the high-resistivity middle layer (30 – 100 kΩ·m) to be frozen till. Figure 5 gives an example (line L_3, distance 140m) where the frozen till was detected at a depth of 0.85 m. The measured temperature, decreasing from +8.7 °C at the surface to +0.8 °C at the bottom of the pit, supports the frost observation (Fig. 5, see also Fig. 8). Another frozen ground observation on this line was made at a distance of 40 m. There, with an iron bar, the frozen till was detected at a depth of 1.5 m. Ten years earlier, in 1994, Hirvas et al. (2000) found frozen till and ice there at a depth of 1.9 m. This observation was made at the same time, the end of July, and at the same place, approximately the middle of line L_3. Because of inaccurate positioning in 1994, the exact location of the drilling site is, however, not known. The maximum difference between the latter observation of this study (40m) and the one made by Hirvas et al. (2000) was estimated to be less than 20 metres.

The high-resistivity layer, reflecting the frozen till, seems to be discontinuous. At the beginning of line L_3, where the line crosses a gabbro outcrop, the resistivity values are 10 – 20 kΩ·m, i.e., markedly lower than those for frozen till. A similar "low-resistivity" feature was also detected in the middle of the line between the distances of 90–110 metres. The reason for the latter low-resistivity feature is not known, but it may be due to a thinner till layer on the more conductive bedrock. The resistivity measured for the gabbro outcrop is close to the resistivity values measured for the samples in the laboratory (Table 2).

Resistivity sounding results from the second test site, situated at 1000–1010m a.s.l. on the western slope of the fell, differ from those measured in the summit

Table 2. Laboratory results for till samples from the summit plateau of the Ridnitšohkka fell (R-1, 1290m a.s.l., middle of line L_3), and from the western slope of the fell (R-2, 1006 m a.s.l., middle of line L_6): Electrical resistivity, pH, cation exchange capacity (CEC), the fines content (∅ <0.06 mm) and clay-fraction content (∅ <0.002 mm) in weight percentages. Temperature +4 °C for the till samples and +20 °C for the gabro samples. The gabbro sample was saturated with tap water (80 Ω·m).

R-1, depth (cm) 1290m a.s.l.	resistivity Ω·m	pH	CEC meq/kg	<0.06mm % p	<0.002mm % p
0 – 3	1460	5.0	0.62	30.8	7.4
3 – 10	4610	5.1	0.65	32	8.5
10 – 15	6750	5.1	0.78	30.1	7.1
15 – 20	5295	5.1	0.76	23.1	5.8
20 – 40	5178	5.1	0.75	25.4	6.2
R-2, depth (cm) 1005m a.s.l.					
0 – 3	2213	4.9	6.59	25.6	6.0
3 – 15	2065	5.0	8.21	25.3	6.7
10 – 20	1974	5.0	8.66	23.2	5.8
30 – 40	2310	4.9	8.55	24.4	6.3
bedrock (gabbro)	Outcrop, beginning of the line L_3, ρ=8-15 kΩ·m. K=420*10-6 SI. D=2.87 kg/m³				

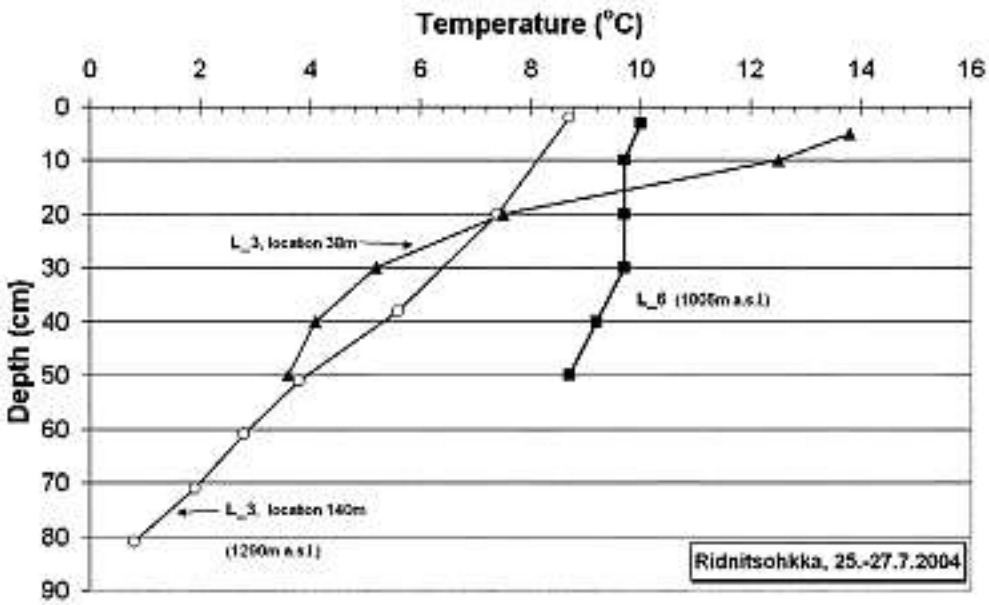


Fig. 8. Temperature-depth profiles from the summit area of the Ridnitšohkka fell (L3, 1290 m a.s.l.) and from the western slope (L6, 1005m a.s.l.).

area. At the lower site the high-resistivity top layer and the very-high-resistivity middle layer are missing. Comparison of the resistivity sections from the top of the fell and from the lower slope leads to the conclusion that the lower site is unfrozen throughout. The fact that the laboratory resistivity values (Table 2) are similar to those of the field values further supports the above conclusion, as well as the temperature distributions in Figure 8.

Electrical resistivity values presented in Table 2 are in agreement with the field data: the mean laboratory values for the till samples are 4700 Ω·m (summit area) and 2100 Ω·m (western slope of the fell), while the comparable field values, i.e., the apparent and inverted resistivities describing the uppermost till layer, are 7000–30000 Ω·m and 1000–10000 Ω·m, respectively. At both sites, the resistivity is higher than the corresponding values in Southern Finland. The cation exchange capacity (CEC) of the till samples, 0.62–0.78 meq/kg at the top and 6.59–8.66 meq/kg on the lower slope of the fell, at least partly explains the difference in the resistivity values. Namely, CEC is a measure of charged ions weakly adsorbed onto mineral and soil particle surfaces and those weakly adsorbed ions, in addition to the free charged particles in the pore water, are largely responsible for the electrical conductivity of sediment.

Conclusions and discussion

Based on the results of the resistivity survey and the other field and laboratory investigations, the till cover in the summit area of the Ridnitšohkka fell seems to be frozen to a depth of several metres. The resistivity sections clearly indicate a one-metre thick unfrozen top layer, the active layer, and a 3–5 metres thick high-resistivity layer beneath it. The high-resistivity layer has been interpreted as frozen till. The high-resistivity layer is underlain by a low-resistivity layer, which has been interpreted as unfrozen bedrock. No drilling data are, however, available and the interpretation of the high-resistivity – low-resistivity boundary is not unambiguous. To produce more reliable estimates of the thickness of the frozen layer, reference data – drillings, refraction seismic or gravity – are needed. Results from the test area situated 300 metres lower on the western slope of the fell strongly suggest unfrozen conditions there.

The contact between the overburden and the bedrock is sometimes difficult to detect with methods based on electrical conductivity. For example, the resistivity of unsaturated coarse sediments can be close to that of bedrock. In this study, the resistivity of frozen till varied from a few tens of thousands of Ohm-metres to about 100 000 Ω·m. The resistivity of unfrozen bedrock lies within the range of frozen till, or lower. The resistivity of frozen bedrock seems to lie within the range of frozen till, or be higher (Table 1). On the other hand, the uppermost layer of the bedrock on the Ridnitšohkka fell is highly fractured and weathered due to intensive freezing and thawing, which decreases both the frozen and unfrozen resistivity values. Whether the bedrock is highly fractured in the till-covered areas is, however, not known. As seen in Figure 5, the

very-high-resistivity top layer describing the frozen till underlies a low-resistivity layer. The resistivity of this deeper layer is between 400 Ω·m – 1000 Ω·m when derived from ground resistivity soundings, and 50 Ω·m – 500 Ω·m when derived from the AEM data (see Fig. 7). These values indicate highly fractured bedrock, or bedrock in which the electrical conductivity is slightly increased due to metallic minerals. Sulphide minerals have been reported to occur only as accessory minerals in the gabbro sills (Sipilä 1992), but near the contact with cumulates, several metres thick layers containing graphite and sulphides have been found (Sipilä 1992). The sills typically dip 20°–50° towards the cumulate unit (Sipilä 1992). The fact that the dips of the electrical conductors in the airborne resistivity sections are very much the same supports the possibility that they originate from bedrock layers having electrically conducting minerals, possibly the same type of graphite and sulpide-bearing layers as detected near the contact.

If the increased conductivity is due to metallic minerals, no evidence is available whether the rock is frozen or not. If, instead, the conductivity of the rock is increased due to fracturing and weathering, the low resistivity refers to unfrozen rock.

The similarity between the results of the airborne and the ground measurements indicates that an airborne electromagnetic survey could be an applicable method for studying the surface parts of permafrost regions.

Acknowledgements

The Lapland Frontier Guard District and Unibase Oy provided important accommodation and logistics support during the spring and summer 2004–2005 fieldwork.

REFERENCES

Andréasson, P.G., Gee, D.G., Whitehouse, M.J. & Schöberg, H. 2003. Subduction-flip during Iapetus Ocean closure and Baltica-Laurentia collision, Scandinavian Caledonides. Terra Nova 15, 362–369.

Hirvas, H., Lintinen, P. & Kosloff, P. 2000. An extensive permanent snowfield and the possible occurrence of permafrost in till in the Ridnitšohkka area, Finnish Lapland. Bulletin of the Geological Society of Finland 72, 47–56.

Hirvas, H., Lintinen, P., Ojala, A.E.K. & Vanhala, H. Geological characteristics of Haltin- Ridnitšohkka region, Enontekiö, Finland. Geological Survey of Finland, Special Paper, in press (this publication).

Hoekstra, P. & McNeill, D. 1973. Electromagnetic probing of permafrost: Proceedings of the 2nd International Conference on Permafrost. National Academy of Sciences, Washington 517–526.

King, L. & Seppälä, M. 1987. Permafrost thickness and distribution in Finnish Lapland – results of geoelectrical soundings. Polarforschung 57, 127–147.

Lehtovaara, J.J. 1995. Kilpisjärven ja Haltin kartta-alueiden kallioperä. Summary: Pre-Quaternary rocks of the Kilpisjärvi and Halti map-sheet areas 1:100 000, Explanation to the Maps of Pre-Quaternary Rocks, Sheet 1823 and 1842, Geological Survey of Finland, 64 p.

Lehtovaara, J.J. 1987. Enontekiön Ridnitsohkkan emäksinen juoniparvi. Summary: Basic dyke swarm at Ridnitsohkka fell in northwesternmost Enontekiö. In: Aro, K. & Laitakari, I. (eds.), Suomen diabaasit. Geological Survey of Finland, Report of Investigation, 199–203. [in Finnish]

Loke, M.H. & Barker, R.D. 1996. Rapid least squares inversion of apparent resistivity pseudosections by a quasi-Newton method. Geophysical Prospecting 44, 131–152.

Olhoeft, G.R. 1978. Electrical properties of permafrost: Proceedings of the 3rd International Conference on Permafrost. National Resources Council, Canada.

Peltoniemi, M. 1982. Characteristics and results of an airborne electromagnetic method of geophysical surveying. Geological Survey of Finland, Bulletin 321, 229 p.

Peltoniemi, M. 1988. Maa- ja kallioperän geofysikaaliset tutkimusmenetelmät. Applied geophysical techniques 515, Espoo: Otakustantamo, 411 p. [in Finnish]

Poikonen, A., Sulkanen, K., Oksama, M. & Suppala, I. 1998. Novel dual frequency fixed wing airborne EM system of Geological Survey of Finland (GTK). Exploration Geophysics 29, 46–51.

Saarelainen, S. 1990. Routavaurioiden korjausrakenteet valtatiellä 21 Peerassa. Mitoitus, koerakentaminen ja seuranta v. 1986–1988. VTT, Research Notes 1102, 62 p. [in Finnish]

Scott, W.J. & Kay, A.E. 1988. Earth resistivities of Canadian soils: Can. Elect Assoc., Montreal. Vol 1, Main Rep., Vol 2., Site data.

Seppälä, M. 1997. Distribution of permafrost in Finland. Bulletin of the Geological society of Finland 69, 87–96.

Sipilä, P. 1992. The Caledonian Halti-Ridnitsohkka igneous complex in Lapland. Geological Survey of Finland, Bulletin 362, 75 p.

Vanhala, H. 2000. Vastusluotaus Peeran palsalla ja Saanatunturilla elokuussa 1999. (Resistivity soundings at the Peeras palsa and the Saana fell in August 1999). Geological Survey of Finland, Report Q 16.1/2000/3, 19 p. [in Finnish]

Vanhala, H., Lintinen, P. & Lehtimäki, J. 2001. Mapping frozen ground using electrical measurements – a case from Peera, a palsa in Finnish Lapland. In: Hill, I. (ed.), 7th Meeting Environmental and Engineering Geophysics, Birmingham, England, September 2nd–6th 2001, proceedings. Lausanne: Environmental and Engineering Geophysical Society, European Section, 70–71.

Vanhala, H. & Lintinen, P. 2002. Test of geophysics for monitoring frozen ground – a case from the southern limit of discontinuous permafrost in Finnish Lapland. In: The Second AMAP International Symposium on Environmental Pollution of the Arctic, Rovaniemi, Finland, October 1–4, 2002, extended abstracts. AMAP Report 2002:2, 3 p.

Quaternary studies in the northern and Arctic regions of Finland
Edited by Antti E.K. Ojala
Geological Survey of Finland, Special Paper 40, 23– 32, 2005.

DATING OF THE HOLOCENE GLACIER VARIATIONS IN THE HALTI-RIDNITŠOHKKA REGION BASED ON DISTAL LACUSTRINE SEDIMENT CORES

by
Antti E.K. Ojala[1)], Samu E. Valpola[2)], Heikki Hirvas[1)], Petri Lintinen[1)], Heikki Vanhala[1)] & Jari Nenonen[3)]

[1)]*Geological Survey of Finland, P.O. Box 96, FIN-02151, Espoo, FINLAND*
[2)]*Department of Geology , P.O. Box 64, FIN-00014, University of Helsinki, FINLAND*
[3)]*Geological Survey of Finland, P.O. Box 1237, FIN-70211, Kuopio, FINLAND*
E-mail: antti.ojala@gtk.fi, samu.valpola@helsinki.fi, heikki.hirvas@gtk.fi, petri.lintinen@gtk.fi, heikki.vanhala@gtk.fi, jari.nenonen@gtk.fi

Ojala, Antti E.K., Valpola, Samu, Hirvas, Heikki, Lintinen, Petri ,Vanhala, Heikki & Nenonen, Jari, 2005. Dating of the Holocene glacier variations in the Halti-Ridnitšohkka region based on distal lacustrine sediment cores. *Geological Survey of Finland, Special Paper 40*, 23–32, 7 figures.
Palaeomagnetic dating of sediment sequences from lakes Pihtsusjärvi and Haltijärvi in NW Finnish Lapland provided a preliminary chronological framework to study the palaeohydrological and glacier history of the Halti Fell area. Both sediment records exhibited a stable primary component of the NRM direction, which allowed us to compare NRM records between the study lakes and the Lake Nautajärvi reference curve chronology. The Lake Pihtsusjärvi sediment sequence was found to cover the last ca. 10 000 years, whereas the Lake Haltijärvi sequence covers only the last ca. 5000 years. The physical properties of these downstream lacustrine sequences provided evidence of post-glacial activity of the Halti Fell ice tongue. The results indicated that the Halti glacier was active soon after the continental ice sheet had retreated from the area at around 10 000-9000 years ago, during which the Govdajohka valley end-moraines were formed. In addition, the bottom age of Lake Haltijärvi sediment core and Lake Pihtsusjärvi sediment stratigraphy indicated minor, short-term neoglacial activity in the Halti Fell vicinity between ca. 3000 BC and the Holocene thermal maximum (ca. 5500-3500 BC) in the Fennoscandian region.

Key words (GeoRef Thesaurus, AGI): glacial geology, glaciers, glaciation, lake sediments, lithostratigraphy, paleomagnetism, geochronology, Holocene, Enontekiö, Finland

Introduction

The potential of using distal (downstream) lacustrine sediment sequences in reconstructing the fluctuations of Holocene glaciers has been recognized in various studies in Sweden and Norway (e.g. Karlén 1976, 1982, Nesje et al. 1994). It is based on the identification and dating of physical characteristics in the sediment stratigraphy, as was carried out by Karlén (1976, 1982) in Sweden. He systematically investigated down-core variations in loss-on-ignition (LOI) and relative X-ray density (i.e. mineral rich vs. organic rich sediment content) in numerous downstream sedimentary basins, and interpreted their variations in terms of glacier retreat and advance. This relationship is not always straightforward, especially

when the glacier is very close to the study lake basin (Karlén, 1982), but Karlén (1976, 1982) and Leonard (1986) have proposed that most mineral matter from erosion (and possibly also washing from the glacier) is transported into distal basins during or soon after the formation of end moraines. In addition, some of fine-grained mineral material may be deposited under the glacier and mobilised during glacier retreat, and a sediment yield may vary considerably between years depending on the spring runoff intensity.

The purpose of the present study was to adopt Karlén's (1976, 1982) approach and investigate glacier variations in the Halti – Ridnitšohkka fells area at the end and after the last regional deglaciation in northern Finland. The so-called Halti glacier has been recognized as the only independent glacier in Finland that has been active since the disappearance of the Scandinavian continental ice sheet at approximately 10 000 years ago (Kujansuu 1967, Hirvas et al. 2000). Evidence of its post-glacial advances is supported by a series of well-defined end-moraines in the Govdajohkka valley a few kilometres south of Halti fell. However, the age of these end-moraines has remained unresolved, as well as their relevance to the Holocene climate history in northern Finland. The present study aimed to settle this matter using downstream lacustrine sediment sequences from lakes Haltijärvi and Pihtsusjärvi. We applied basic methods for recording sediment physical parameters (LOI, mineral magnetic parameters, and detailed sediment stratigraphy dated with a palaeomagnetic dating method) aiming to reconstruct the glacial and post-glacial palaeohydrological history of the Halti area. We suggest that evidence of advances and retreats of the Halti glacier may be obtained from the continuous sediment records of these lakes, which are then combined with previous moraine stratigraphic investigations by Hirvas (1968) and Hirvas et al. (2000).

Study area

The region investigated is located in NW Finnish Lapland in the municipality of Enontekiö, close to Swedish border and only some 40 km from the Arctic Ocean (N69°15', E21°15') (Fig. 1). It is characterized by high relative differences in altitude, including the highest summits in Finland, Halti (1328 m) and Ridnitšohkka (1315 m), and broad valleys in between high-peaked fells. Numerous lakes and rivers are situated in these valleys, including the present study lakes Pihtsusjärvi and Haltijärvi. The bedrock of the area consists of Caledonian amphibolite, muscovite gneiss, and garnetiferous mica gneiss (Nabar sheet) overthrusted on the Archaean basement. In Halti and Ridnitšohkka fells this unit is overlain by dunite and gabbro (Korrovarri sheet), respectively (Lehtovaara 1994, 1995).

The deglaciation of NW Finnish Lapland began about 11 500 years ago (Kujansuu 1967, Johansson & Kujansuu 2005). Unlike in southern Finland, where the continental ice sheet terminated at the Baltic basin and diminished quickly through a calving process, deglaciation in NW Finnish Lapland was substantially slower and occurred mainly as gradual thinning of the ice sheet. The highest peaks of the fell area were first exposed from underneath the ice cover as nunataks, whereas valleys were covered with separate bodies of dead ice as deglaciation proceeded. It is evident that some of these lobes of ice and cirque-like basins have functioned independently after separation from the continental ice sheet (Kujansuu 1967). Halti Fell ice tongue is probably the best representative of a post-glacial valley glacier in Finland. Arching end-moraines in the Govdajohkka valley, which also contain dunite boulders that are exclusively found from Halti-type, suggest that the Halti Fell ice tongue probably advanced and retreated on several occasions after the retreat of the continental ice sheet from the area. It extended four to five kilometres south from Halti Fell at the maximum stage. The stratigraphy, number and locations of end moraines set up by these glacier fluctuations have been further studied by Hirvas (1968) and R. Koskinen (personal communication, 2005). There are 6 continuous and larger scale moraine arches in the area, and 7 to 9 scattered minor features based on fieldwork and aerial photographic interpretations.

Lake Pihtsusjärvi has an excellent position to study hydrological and erosional changes caused by movements of the valley glacier of Halti in post-glacial times. The lake has two separate parts, a southern basin and northern basin, which have maximum depths of 13.6 and 20.6 metres, respectively. The southern basin of the lake is of particular interest in this study, because it is located some 2 kilometres south of the end moraines and there is a small creek running into the basin from the Govddajohka valley. It basically traps all the material that is and has been transported southwards via Govddajohka creek during the last ca. 10 000 years. The other site in the present study, Lake Haltijärvi, is located 5 km north of Lake Pihtsusjärvi and is substantially smaller and shallower with a maximum depth of 6.6 m. It is, however, located directly below the fell Halti and has probably been the last area exposed from underneath the valley glacier of Halti. Lake Haltijärvi lies at an altitude of 922 m a.s.l., whereas Lake Pihtsusjärvi is almost 200 metres lower (739 m a.s.l.). The shorelines of both lakes are very

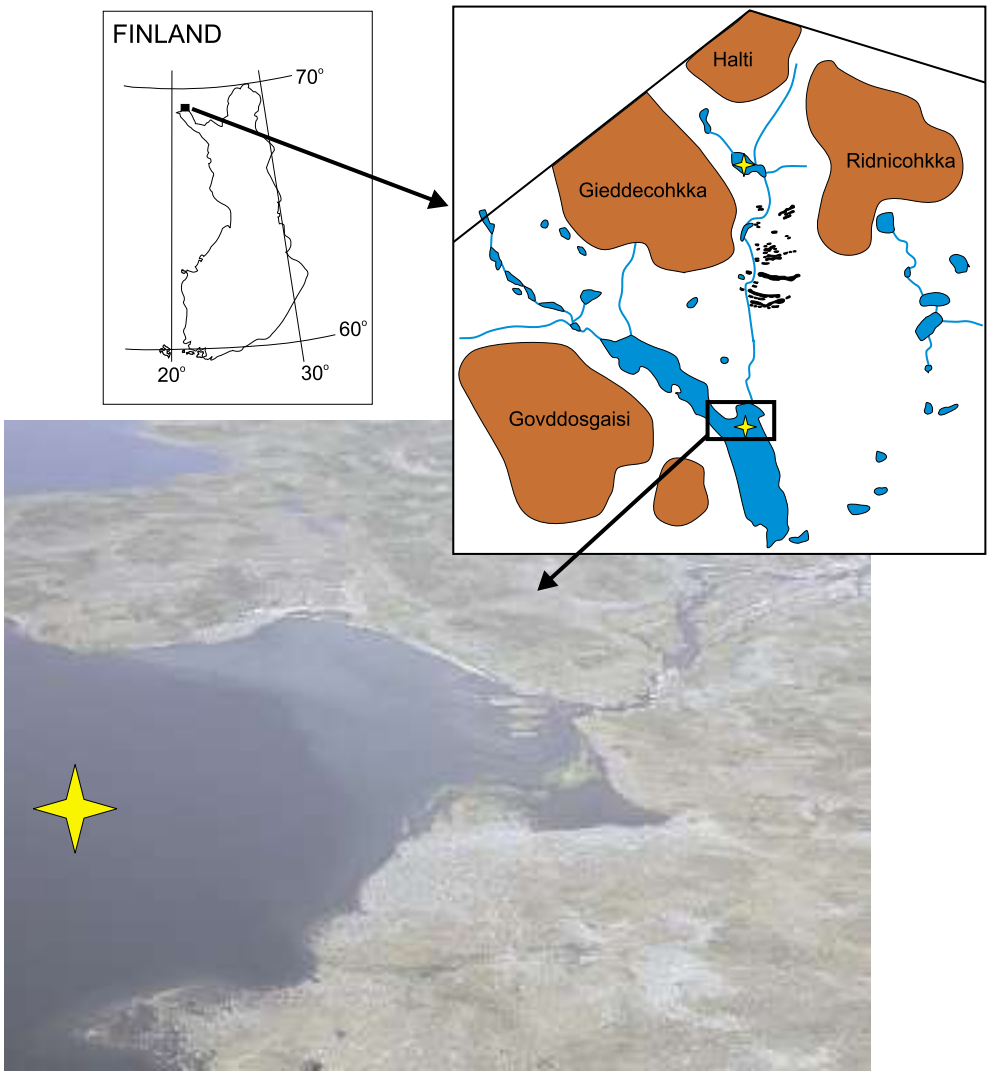


Fig 1. Map showing the location of the study area with coring sites in Lakes Pihtsusjärvi and Haltijärvi marked with stars. Curved black lines in the upper-right of the picture represent Govdajohkka valley end-moraine formations (as investigated by R. Koskinen, personal communication). (Photo: A.E.K. Ojala, 2004)

rocky and the catchments are covered by grassland and shrub-type vegetation.

The study area is located in the hemi-arctic region between the North Atlantic Ocean climate and the Eurasian continental climate. The mean annual temperature is around -3 °C, the warmest month being July (ca. +10 °C) and the coldest January or February (ca. -15 °C). The mean annual precipitation is around 500 mm, of which more than half falls as snow. The winter snow cover usually appears in October and disappears in early June (Alalammi 1987). The only area that has been reported to maintain a year-round permanent snow cover in Finland is located in the eastern corner of the study area, on the eastern slope of Ridnitšohkka Fell (Hirvas et al. 2000). The study lakes are ice-free for about 3 months during the summer.

Materials and methods

Results of this study are based on palaeolimnological and sedimentological investigations of the sediment sequences retrieved from lakes Pihtsusjärvi and Haltijärvi. Coring of the lakes was carried out with a light piston corer operated with a series of 2-metre-long rigid aluminium rods in April 2004 (Fig. 2). Coring was conducted at the deepest point of Lake Haltijärvi (6.6 m), which is situated some tens of metres on the SW side of the lake centre (Finnish grid: 1510919/7689120). Altogether, 3 parallel sets of cores were taken, Ha1 (115 cm), Ha2 (134.5 cm), and Ha3 (120 cm), which covered the maximum thickness of sediment that could be penetrated with light-weight equipment. In Lake Pihtsusjärvi, the coring was tar-

geted at the north side of the southern basin (Finnish grid: 1511175/7683549) (Fig. 1). Two parallel cores, Pi1 (189 cm) and Pi2 (196 cm), were taken from the depth of 13.7 metres, just south of the littoral zone delta seen in Figure 1. Samples were prevented from freezing and stored at cold room temperature (+4 °C).

Lengthwise split sediment cores were described according to the Troels-Smith classification (Troels-Smith 1955), paying special attention to the presence and location of characteristic clay layers, which were used as "marker horizons" to correlate between parallel cores. Magnetic susceptibility (κ) was then measured at 0.5 cm resolution from the cleaned and levelled fresh sediment surfaces using a Bartington MS2E1 sensor. This is a measure of the concentration and grain-size of magnetic minerals in the sediment and a potential indicator of erosion-related processes within the catchment (e.g. Snowball & Sandgren 2001). Loss on ignition (LOI) was determined at 1 cm resolution by keeping the oven-dried samples (105 °C, overnight) in a furnace at 550 °C for 2 h.

Cores Ha2 and Pi2 were continuously sampled for palaeo- and mineral magnetic measurements at 2.5-cm intervals using 7 cm^3 polystyrene cubes. The cubes were oriented according to the sample tube and pressed perpendicularly into the levelled sediment surface. Natural remanent magnetization (NRM), including declination, inclination and intensity, was measured with a 2G-Enterprises SRM-755R tri-axial SQUID magnetometer at the Geological Survey of Finland's palaeomagnetic laboratory. First, pilot samples from different stratigraphical units of both lakes were selected for stepwise AF demagnetization (0 to 120 mT peak AF), aiming to test the stability of the NRM and the carrier of the remanence. Second, all sample cubes were measured for NRM, then demagnetized along three axes with a 20 mT peak alternating field (AF) and measured again. Following these measurements, samples were subjected to anhysteretic remanent magnetization (ARM), isothermal remanent magnetization (IRM) and saturation isothermal remanent magnetization (SIRM) to determine the carrier of the remanence and to study stratigraphical variations in magnetic mineral concentration and grain sizes (e.g. Snowball & Sandgren 2001). ARM was induced to all samples with a biasing direct field of 0.05 mT superimposed on a peak alternating field of 100 mT. AF demagnetized samples were then submitted to IRM acquisition with a Molspin pulse magnetizer by increasing the strength of the magnetic field stepwise from 25 to 1500 mT. The remaining samples were exposed to a 1000 mT maximum field, which was considered to represent a point of saturation (SIRM) in most samples. IRM and SIRM were measured with a Molspin spinner magnetometer. In addition, a backfield IRM at 100 mT was induced and measured again with a Molspin spinner magnetometer. The S-ratio was determined as $IRM_{-100mT}/SIRM_{1000mT}$ (Stober & Thompson 1979).

Fig. 2. Coring of Lake Pihtsusjärvi with a light piston corer in April 2004. (Photo: A.E.K. Ojala, 2004)

Results and discussion

Correlation and dating of the sediment cores

Lengths of the sediment cores from Lakes Haltijärvi and Pihtsusjärvi, respectively, ranged from 115 to 134.5 cm and from 189 to 196 cm. In both lakes, stratigraphic characteristics as well as susceptibility logs indicated that parallel cores were very much alike, allowing a reliable core-to-core correlation within both basins. This indicates that there are stable and continuous sedimentary environments in the coring locations, and that the coring itself proceeded undisturbed. Both facts are fundamentally important for the study of sediment physical parameters and the dating of sequences. In addition, mineral magnetic parameters and LOI correlated well with the visual sediment stratigraphy.

Stepwise IRM acquisition and AF demagnetization

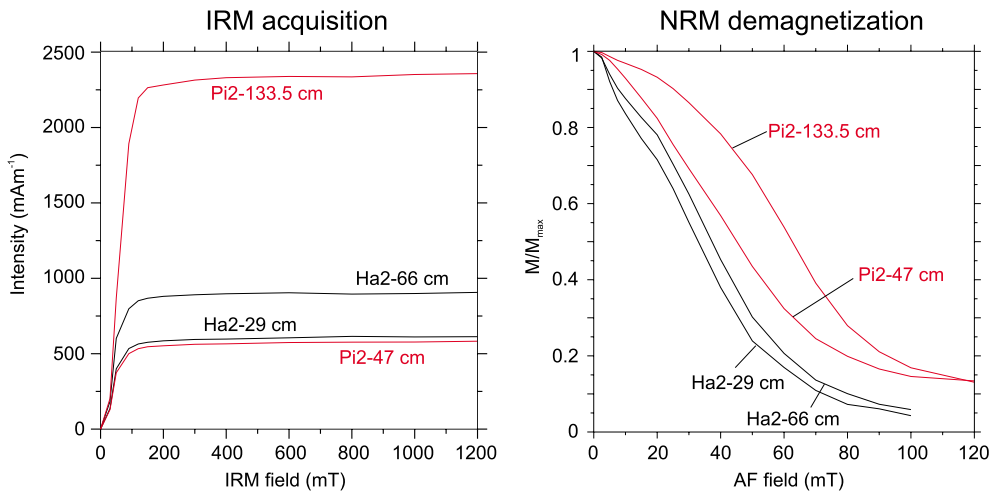


Fig. 3. IRM acquisition and NRM demagnetization behaviour of the pilot samples Pi2-47 and Pi2-133.5 (from Lake Pihtsusjärvi), and Ha2-29 and Ha2-66 (from Lake Haltijärvi) indicate that fine-grain magnetite is the major carrier of the remanence in Lake Haltijärvi, whereas Lake Pihtsusjärvi sediments contain some amount of "harder" magnetic mineral or minerals. M/Mmax is the normalized intensity and MDF represents the median destructive field of NRM.

of NRM behaviour of the pilot samples are plotted in Figure 3. These curves indicate that ferrimagnetic minerals (most likely magnetite) carry approximately 95% of the remanence in Lake Haltijärvi sediments (Thompson & Oldfield 1986). Pilot samples Ha2–29 and Ha2–66 are largely demagnetized in the AF field of 100 mT, and have median destructive field (MDF) values of 33 and 38 mT, indicating magnetic behaviour typical of fine-grained (SD and PSD) magnetite (King et al. 1982, Thompson 1986). Samples were also saturated (SIRM) in fields less than 200 mT, which also indicates magnetite dominance as a magnetic mineral (King et al. 1982, Thompson 1986). Pilot samples Pi2–47 and Pi2–133.5 from Lake Pihtsusjärvi, however, remain unsaturated in the highest IRM field (1200 mT) and are more resistant to high AF-demagnetization. They also have MDF values of 57 and 73 mT. In addition to magnetite, Lake Pihtsusjärvi sediments probably contain some canted anti-ferrimagnetic mineral or minerals (e.g. hematite) that are carrying approximately 15% of the remanence (Thompson 1986).

The stability of the NRM direction was investigated with Zirderveld orthogonal vectors plotted for pilot samples based on AF demagnetization spectra. All pilot samples in both lakes exhibited a stable primary component of the NRM direction, as given in an example from Lake Pihtsusjärvi cube Pi2-133.5 (Fig. 4). This allowed us to compare NRM records between the study lakes and the Lake Nautajärvi reference curve (Ojala & Tiljander 2003) for constructing a palaeomagnetic chronology.

The chronology of the Lake Nautajärvi palaeosecular variation (PSV) reference curve (i.e. master curve) is based on the counting of lacustrine varves, and it has a cumulative dating error of less than 1% for the entire 10 000-year sequence (Ojala & Tiljander 2003). PSV curves for the sediment sequences of lakes Haltijärvi and Pihtsusjärvi are plotted and compared with the Nautajärvi master curve in Figures 5a and 5b. Even though their resolutions differ from the reference curve resolution, they show many of the dominant swings in NRM direction previously identified in Finland, Sweden and the British Isles (Turner & Thompson 1981, Snowball & Sandgren 2002, Ojala & Tiljander 2003). According to NRM measurements, Lake Haltijärvi sediment deposits appear to cover only the last ca. 5000 years, whereas Lake Pihtsusjärvi sediment deposits extend back about 10 000 years. These results are based on NRM investigations of single cores,

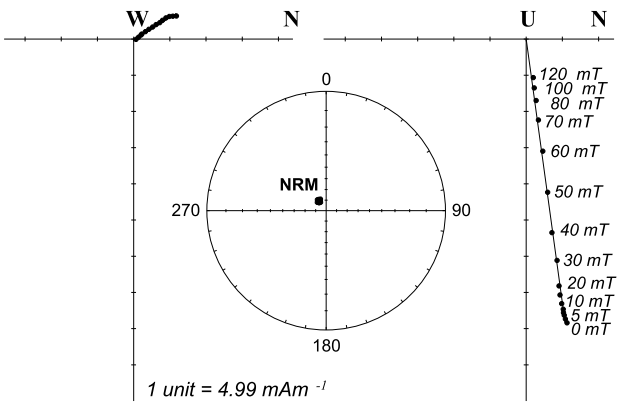


Fig. 4. Orthogonal Zijderveld projections of the stepwise AF demagnetization of NRM calculated for the pilot sample Pi2-133.5. The magnetic signal is dominated by a strong and stable primary component of the magnetic direction.

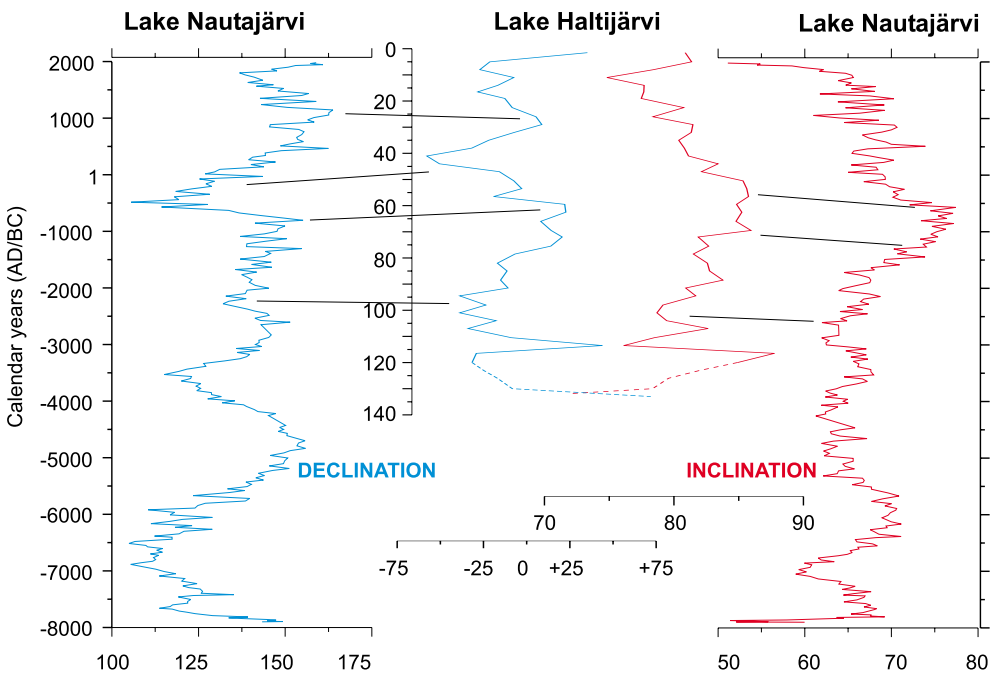


Fig. 5a. Palaeosecular variation curves, declination and inclination, of Lake Haltijärvi correlated with the varve-dated (AD/BC) reference curves from Lake Nautajärvi, central Finland (Ojala & Tiljander 2003).

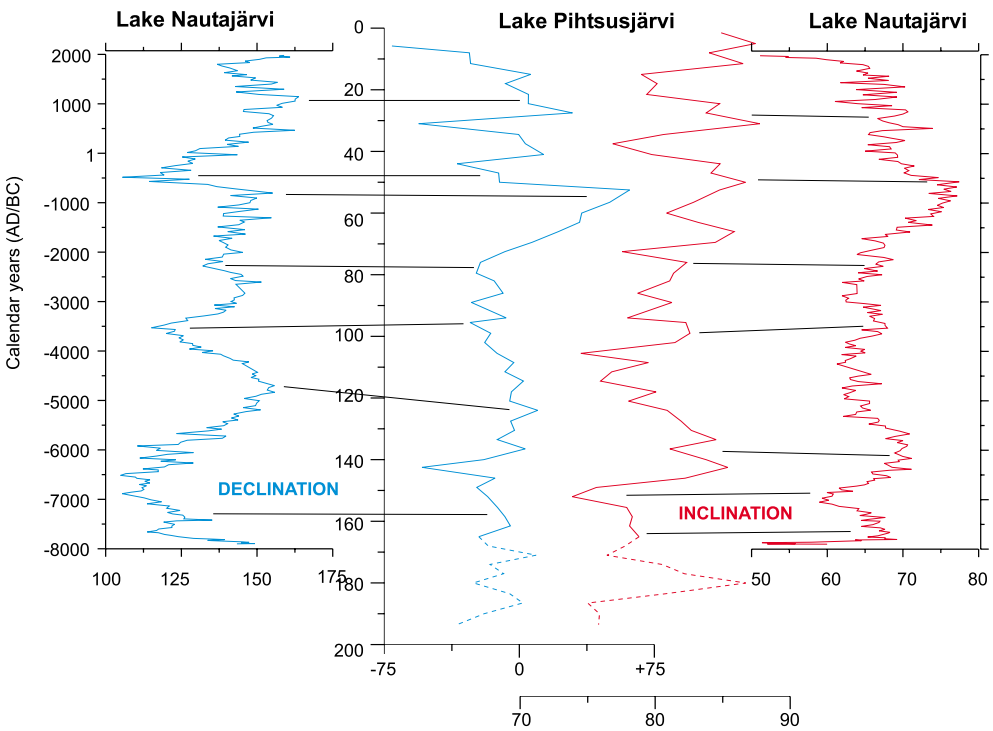


Fig. 5b. Palaeosecular variation curves, declination and inclination, of Lake Pihtsusjärvi correlated with the varve-dated (AD/BC) reference curves from Lake Nautajärvi, central Finland (Ojala & Tiljander 2003).

however, and must be taken as preliminary at the moment. The final age-depth transformations for lakes Haltijärvi and Pihtsusjärvi will be based on repeated NRM measurements on a new set of parallel cores from both lakes. These cores were retrieved using a heavy piston corer in April 2005.

Sediment lithostratigraphy of the Lake Haltijärvi sequence

The sediment stratigraphy of Lake Haltijärvi can be divided into 3 lithostratigraphic units (Fig. 6). The topmost sediment (0–10 cm) is composed of light

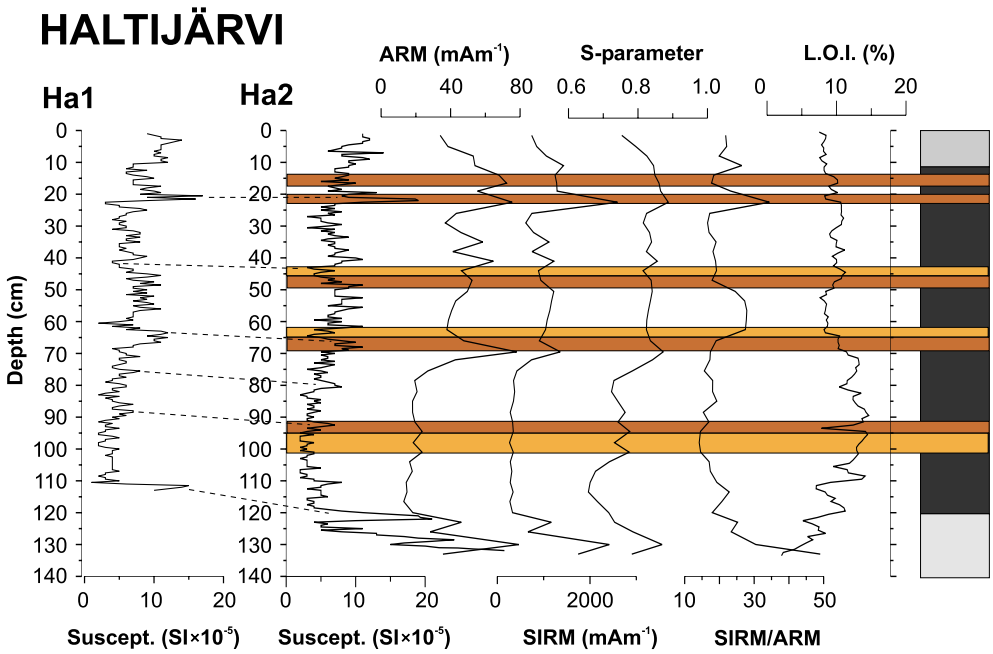


Fig. 6. Lake Haltijärvi sediment lithostratigraphy, LOI and mineral magnetic parameters for the 140 cm long sequence Ha2. The left-hand side susceptibility curve from the core Ha1 shows the excellent correspondence between parallel cores within the lake.

homogeneous gyttja that has a high water content because it has not yet been compacted. Susceptibility is decreasing downwards in this unit, while ARM and SIRM are increasing, suggesting a decrease in the concentration of magnetic minerals but an increase in their grain size.

The middle unit of the Lake Haltijärvi sediment sequence (10–120 cm) possesses the most interesting stratigraphic fluctuations. Four depths (ca. 15–20, 45–50, 65–70, and 90–100 cm) are characterized by approximately 5 cm thick sections of darker and lighter bands in the sediment, which can also be detected in mineral magnetic parameters and LOI. Often, the darker band lies below the lighter one, and there is a sharp contact in between. Preliminary investigations suggest that the lighter bands are composed of clayed mineral particles, whereas the darker ones contain sulphic material and possibly darker minerals. Although the chemical and mineralogical analysis of these layers is still ongoing, one could suggest that there is a sedimentological forcing factor that is driving their rhythmic fluctuations. This phenomenon is likely to be related to catchment erosion and an allochthonous sediment yield into the basin. Below the depth of 120 cm the sediment is composed of gyttja clay and clay that has no clear structures. The higher concentration of magnetic minerals indicate a different sedimentological environment from that above, possibly representing the early stages of the basin having a vast amount of redeposited mineral material.

Preliminary results from palaeomagnetic dating of the Lake Haltijärvi sediment sequence raise an interesting question about the age of the basin. As approximately equally long sediment cores (ca. 130 cm) were retrieved with the heavy piston corer in 2005 compared with spring 2004, it is likely that the cores presented here cover the entire sediment sequence deposited in Lake Haltijärvi. If the age, ca. 5000 years, for Lake Haltijärvi turns out to be of good accuracy in replicated NRM measurements, what is the reason for such a late development of the lake? Looking at other palaeoclimate records in Sweden, Finland and Norway, it would appear very unlikely that the valley glacier of Halti would have survived un-melted during the so-called Holocene thermal optimum (HTM) in Fennoscandia (ca. 5500–3500 BC) (Koç & Jansen 1994, Seppä & Birks 2001, Snowball et al. 2004). One possibility is that that the remains of the old glacier were covered by a vast amount of melt-out till during the last stages of the glacier's evolution, and then slowly melted during the next thousands of years, finally forming the Lake Haltijärvi basin. More likely, there was minor neoglacial activity in the Halti Fell vicinity between ca. 3000 BC and the HTM, which allowed the development of a local glacier or a round-the-year remnant of snow and ice in the place where Lake Haltijärvi is presently located. Such a mass of snow and ice then slowly scraped earlier sediment deposits from Lake Haltijärvi, but did not have enough critical mass to advance and form visible end-moraines. According

to the Lake Haltijärvi results, the neoglacial activity lasted for a short period of less than a thousand years between approximately 3500 BC and 3000 BC, which corresponds well to other studies of Holocene glacier variation in Norway and Sweden (e.g. Karlén 1988, Matthews et al. 2000). Karlén (1988) reported a series of mid-Holocene glacier advances in northern Sweden between 5100 and 4500 uncalibrated radiocarbon years ago, whereas Matthews et al. (2000) suggested the existence of small glaciers in southern Norway post-5200 cal. BP. In addition, Hyvärinen & Alhonen (1994) have reported that this period (ca. 6000–5000 years ago) was characterized by very low (up to 4 metres lower) water levels in lakes in western Finnish Lapland due to severe dryness. As a consequence, considerable erosion and redeposition of material may have occurred in a lake such as Haltijärvi during and after these fluctuations.

The only insitu deposits found in Lake Haltijärvi have been formed during the last approximately 5000 years. The rhythmic sediment stratigraphy and stability of NRM data suggest continuous and stable sedimentation in Lake Haltijärvi over this period. This would mean that no local glacier has appeared in the Lake Halti vicinity during the so-called Little Ice Age, even if many other Fennoscandian glacier advances seem to be associated with this historically known colder period. However, rhythmic stratigraphic fluctuations in the Lake Haltijärvi sediment sequences may be related to the Late-Holocene climate fluctuations recorded in numerous local glaciers in Norway and Sweden (e.g. Karlén 1988, Matthews et al. 2000).

Sediment lithostratigraphy of the Lake Pihtsusjärvi sequence

The sediment stratigraphy of Lake Pihtsusjärvi is presented in Figure 7. The preliminary dating results for the base of this sequence, ca. 10 000 years, corresponds well with previous deglaciation history investigations from the area (e.g. Johansson & Kujansuu 2005). The sequence can be divided into three lithostratigraphical main units, of which the uppermost unit (0–10 cm) is very similar to the uppermost unit of the Lake Haltijärvi sequence. However, there are no detectable changes in either the concentration or grain size of the magnetic mineral, but LOI has a clear declining trend between 0 and 20 cm.

The middle unit of the Lake Pihtsusjärvi sequence (10–180 cm) could have been further divided into several subunits based on the physical properties of the sediment. There is considerable fluctuation in the measured parameters within this unit, which are seen in susceptibility logs, in particular. The susceptibility maximum peak at the depth of 93 cm has an age of ca. 5000 years and is probably related to neoglacial activity detected in the Lake Haltijärvi vicinity. The concentration of magnetic minerals increases with depth between 140 and 180 cm, being three times higher in the lower part of the unit and having a high-frequency variability. By contrast, the S-parameter and SIRM/ARM ratio remain unchanged, suggesting that magnetic mineralogy and grain size is not changing. Susceptibility maximum peaks in the lower parts of this unit are worth special attention, because they

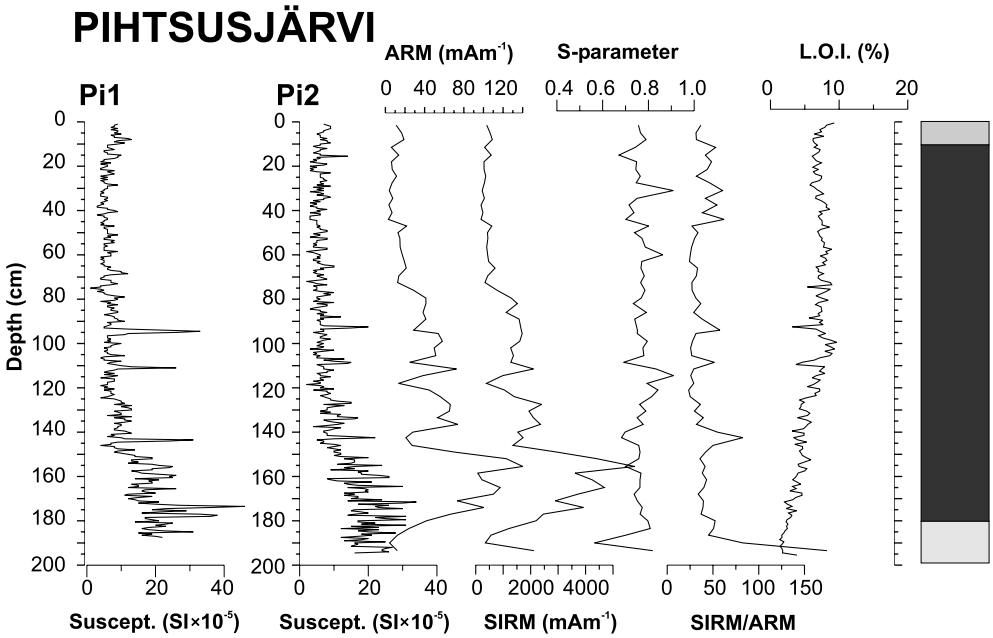


Fig. 7. Lake Pihtsusjärvi sediment lithostratigraphy, LOI and mineral magnetic parameters for the 140 cm long sequence Pi2.

probably reflect advances and retreats of the Halti Valley glacier and the formation of the Govdajohka end-moraines. There are 6 maximum peaks in the susceptibility curve of the Lake Pihtsusjärvi sediment sequence Pi2 between 140 and 180 cm, which can be detected visually as 0.5–20 mm thick layers of mineral matter. This number, plus some minor peaks, agrees well with the number of continuous (plus some scattered) end-moraine formations appearing in the valley as investigated by R. Koskinen (personal communication, March 2005) from aerial photographs. We suggest that these mineral matter laminae originate from erosion during or soon after the formation of end moraines, and possibly also from washing from the glacier during its retreat. This would mean that the Halti glacier was active soon after the continental glacier had retreated from the area, at around 10 000–9000 years ago (8000–7000 BC), and the Govdajohka valley end-moraines were formed during this ca. 1000-year period. To further support this hypothesis, we investigated the size, structure and exact locations of these terminal moraines with ground penetrating radar during the field season in April 2005, but the work is still in progress.

The base of the core, the lowermost unit between 180 and 196 cm, is composed of massive clay (180–190 cm) representing a local short-term glacial lake and classical varved clay deposits (190–196 cm) as a consequence of continental ice sheet retreat from the area towards the south. It is likely that there was no large-scale ice lake in the area, but that meltwaters were instead temporarily blocked in front of the continental ice that was retreating downhill (Johansson & Kujansuu 2005).

Conclusions

The first stage of palaeomagnetic investigations indicated that it is possible to successfully apply palaeomagnetic dating to Lake Pihtsusjärvi and Haltijärvi sediment sequences and then compare their results with the Nautajärvi master curve chronology. Both lakes exhibited a strong and stable primary component of NRM, and contained fine-grained magnetite as the dominant magnetic mineral. Palaeomagnetic dating results and detailed age-depth curves will be finalized after the analysis of parallel sediment cores that were taken with a heavy piston corer in spring 2005.

The preliminary age of the Lake Haltijärvi basin (ca. 5000 years) turned out to be the most surprising result of the present study and was interpreted as an indication of minor neoglacial activity between the HTM and ca. 3000 BC. Lake Pihtsusjärvi sediment deposits were found to cover the last ca. 10 000 years.

This study demonstrated that information can be gained on the activity of a local glacier such as the Halti valley glacier by using downstream sedimentary basins as archives of hydrological changes. Advances and retreats of the Halti glacier were visible in Lake Pihtsusjärvi deposits as layers of mineral material. According to the preliminary results, the Halti glacier was active soon after the continental ice sheet had retreated from the area at around 10 000–9000 years ago, forming the Govdajohka valley end-moraines.

Acknowledgements

The authors are grateful to reviewers M. Saarnisto and M. Eronen for perceptive comments and suggestions, and to R. Koskinen who examined the Govdajohka valley end-moraines. The Lapland Frontier Guard District provided important accommodation and logistics support during the spring and summer 2004–2005 fieldwork.

REFERENCES

Alalammi, P. (ed.) 1987. Atlas of Finland, 131: Climate. Helsinki: National Board of Survey, Geographical Society of Finland, 31 p.

Hirvas, H. 1968. Haltin alueen kvartäärisestä kehityksestä. Pro Gradu –tutkielma. Helsingin yliopisto, Geologian ja paleontologian laitos, 49 p. [in Finnish]

Hirvas, H., Lintinen, P. & Kosloff, P. 2000. An extensive permanent snowfield and the possible occurrence of permafrost in till in the Ridnitšohkka area, Finnish Lapland. Bulletin of the Geological Society of Finland 72, 47–56.

Hyvärinen, H. & Alhonen, P. 1994. Holocene lake-level changes in the Fennoscandian tree-line region, western Finnish Lapland: diatom and cladoceran evidence. The Holocene 4, 251–258.

Johansson, P. & Kujansuu, R. (eds.) 2005. Pohjois-Suomen maaperä: maaperäkarttojen 1:400 000 selitys (Quaternary deposits of Northern Finland – Explanation to the maps of Quaternary deposits 1:400 000). Espoo: Geologian tutkimuskeskus, 236 p. [in Finnish]

Karlén, W. 1976. Lacustrine sediments and tree-limit variations as indicators of Holocene climatic fluctuations in Lappland, Northern Sweden. Geografiska Annaler 58A, 1–34.

Karlén, W. 1982. Holocene glacier fluctuations in Scandinavia. Striae 18, 26–34.

Karlén, W. 1988. Scandinavian glacial and climate fluctuations during the Holocene. Quaternary Science Reviews 7, 199–209.

King, J., Banerjee, K.S., Marvin, S. & Özdemir, Ö. 1982. A comparision of different magnetic methods for determining the relative grain size of magnetite in natural materials: some results from lake sediments. Earth and Planetary Science Letters 59, 404–419.

Koç, N. & Jansen, E. 1994. Response of the high-latitude Northern Hemisphere to orbital climate forcing: Evidence from the Nordic Seas. Geology 22, 523–526.

Kujansuu, R. 1967. On the deglaciation of western Finnish

Lapland. Bulletin de la Commission Géologique de Finlande 232, 98 p.

Lehtovaara, J.J. 1994. Geological Map of Finland, Pre-Quaternary rocks, 1:100 000, sheet 1842 (Halti).

Lehtovaara, J.J. 1995. Kilpisjärven ja Haltin kartta-alueiden kallioperä. ("Pre-Quaternary rocks of the Kilpisjärvi and Halti map-sheet areas"). Geological Survey of Finland, Explanation to the maps of Pre-Quaternary rocks, sheets 1832 and 1842, 64 p. [in Finnish]

Leonard, E.M. 1986. Varve studies at Hector Lake, Alberta, Canada, and relationship between glacial activity and sedimentation. Quaternary Research 25, 199–214.

Matthews, J.A., Dahl, S.O., Nesje, A., Berrisford, M.S. & Andersson, C. 2000. Holocene glacier variations in central Jotunheimen, southern Norway based on distal glaciolacustrine sediment cores. Quaternary Science Reviews 19, 1625–1647.

Nesje, A., Dahl, S.O., Løvlie, R. & Sulebak, J.R. 1994. Holocene glacier acyivity at the southestern part of Hardangerjøkulen, central-southern Norway: evidence from lacustrine sediments. The Holocene 4, 377–382.

Ojala, A.E.K. & Tiljander, M. 2003. Testing then fidelity of-sediment chronology: comparison of varve and paleomagnetic results from Holocene lake sediments from central Finland. Quaternary Science Reviews 22, 1787–1803.

Seppä, H. & Birks, H.J.B. 2001. July mean temperature and annual precipitation trends during the Holocene in the Fennoscandian tree-line area: pollen-based climate reconstructions. The Holocene 11, 527–539.

Snowball, I. & Sandgren, P. 2001. Application of mineral magnetic techniques to paleolimnology. In: Last, W.M. & Smol, J.P. (eds.), Developments in Paleoenvironmental Research. Tracking Environmental Change Using Lake Sediments, Volume 2: Physical and Geochemical Methods. Dordrecht: Kluwer Academic Publisher, 217–237.

Snowball, I.F. & Sandgren, P. 2002. Geomagnetic field variations in northern Sweden during the Holocene quantified from varved lake sediments and their implications for cosmogenic nuclide production rates. The Holocene 12, 517–530.

Snowball, I., Korhola, A., Briffa, K. & Koç, N. 2004. Holocene climate dynamics in Fennoscandia and the North Atlantic. In: Battarbee, R.W. et al. (eds.), Past Climate variability through Europe and Africa. Dordrecht: Springer, 465–494.

Stober, J.C. & Thompson, R. 1979. An investigation into the source of magnetic minerals in some Finnish lake sediments. Earth and Planetary Science Letters 45, 464–474.

Thompson, R. 1986. Modelling magnetization data using SIMPLEX. Physics of the Earth and Planetary Interiors 42, 113–127.

Thompson, R. & Oldfield, F. 1986. Environmental magnetism. London: Allen and Unwin, 227 p.

Troels-Smith, J. 1955. Characterisation of unconsolidated sediments. Geological Survey of Denmark, IV, 3, 73 p.

Turner, G.M. & Thompson, R. 1981. Lake sediment record of the geomagnetic secular variation in Britain during Holocene times. Geophysical Journal of the Royal Astronomical Society 65, 703–725.

Quaternary studies in the northern and Arctic regions of Finland
Edited by Antti E.K. Ojala
Geological Survey of Finland, Special Paper 40, 33– 39, 2005.

MELTWATER CANYON LAKES (SAIVOS) IN WESTERN FINNISH LAPLAND

by
Peter Johansson

Geological Survey of Finland, P.O. Box 77, FIN-96101, Rovaniemi, FINLAND
E-mail: peter.johansson@gtk.fi

Johansson, Peter, 2005. Meltwater canyon lakes (saivos) in western Finnish Lapland *Geological Survey of Finland, Special Paper 40*, 33–39, 5 figures.

In western Finnish Lapland there are circular and hollow-like canyon lakes known as "saivos", the most famous of which are Pakasaivo and Äkäsaivo. The water depth of Pakasaivo is 60 metres, and with the steep cliffs rising above the water level, the total depth of the hollow is over 100 metres. The saivos are polygenetic. They are situated in fracture zones, where the bedrock was already deeply eroded in preglacial time. During the glaciations, the glacial erosion also played a role in the formation of the canyons, but the most important factor by far was the erosion caused by the whirling meltwater stream. The hollow of Pakasaivo was formed in the initial phase of deglaciation, when meltwater erosion was the main subglacial process. The erosional stage was followed by deposition, when the Pakajärvi esker accumulated at the base of the meltwater conduit. Äkäsaivo probably deepened as a subglacial meltwater stream was flowing through it during deglaciation that predated the Late Weichselian glaciation. The till-covered esker sequences in the area are probably related to the same deglaciation. During the final phase of the last deglaciation, waters discharged from the ice-dammed lake covering the Muoniojoki river valley, first through Äkäsaivo, and then shifting to Pakasaivo. The waters that discharged from the ice lake only completed the subglacial meltwater erosion, but not significantly contributing to the formation of the saivos.

Key words (GeoRef Thesaurus, AGI): lakes, canyons, fracture zones, erosion, water erosion, subglacial environment, meltwater channels, Muonio, Finland

Introduction

Saivos are round lakes or ponds with clear water, typically of a considerable depth and generally occurring in connection with glaciofluvial systems. According to Sámi legends (Paulaharju 1922), only lakes with no distinct surface flow or discharge streams can be considered saivos proper. They receive their water from underwater springs and water also flows out of them into bedrock fractures. The saivos are impressive erosional features, often bordered by block-laden steep shores or almost vertical rock walls. Geologically, the saivos are not a uniform type of feature, but at one or more stages their formation has been influenced by the meltwater action of the continental ice sheet.

In the western part of northern Finland, in the vicinity of the Ylläs fell range, there are two well-known saivos, Pakasaivo and Äkäsaivo. They are located about 120 km north of the Arctic Circle. It is an almost uninhabited wilderness region between the western Lapland mountain region and Muonionjoki, the river on the border between Finland and Sweden

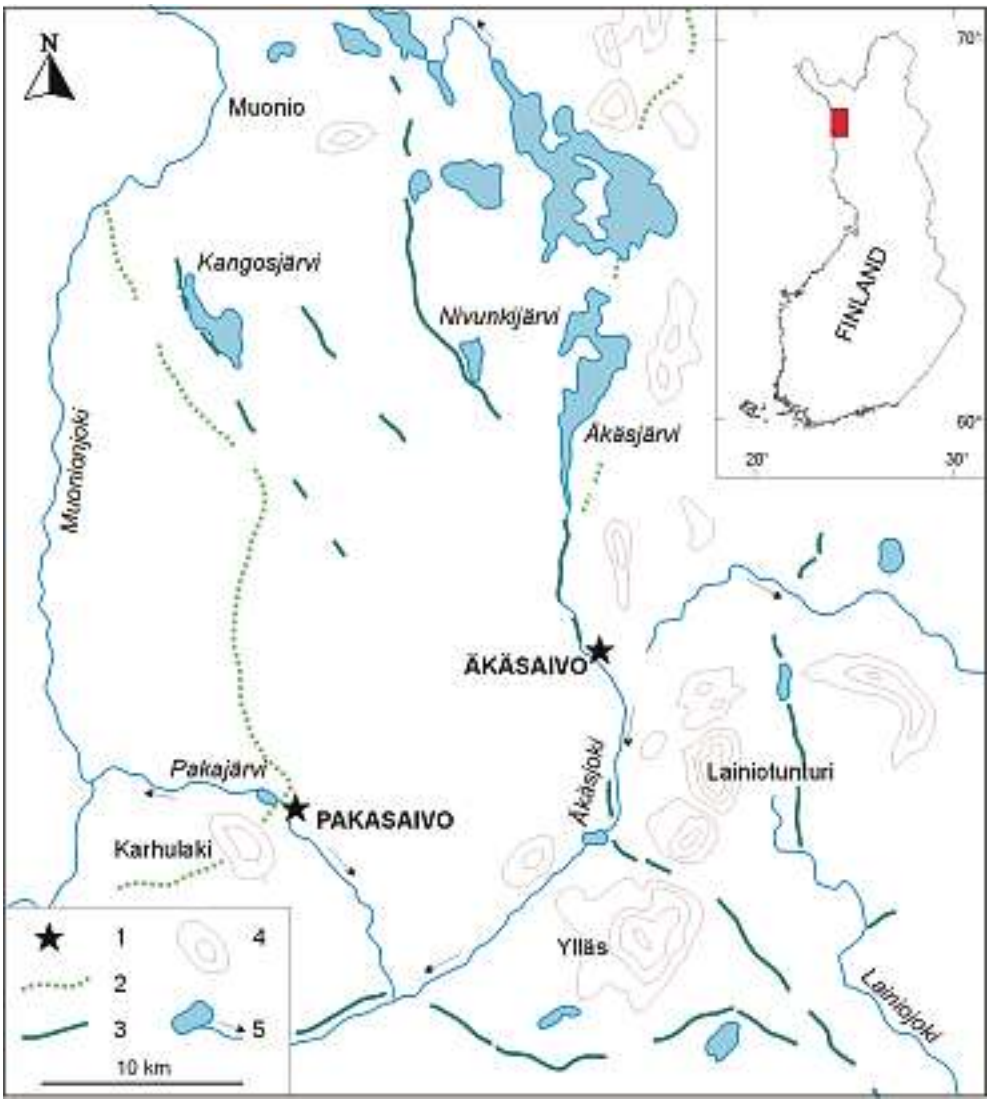


Fig. 1. Location of the study area. 1 = Saivo lakes, 2 = Late Weichselian esker chain, 3 = Early Weichselian esker chain, 4 = mountain or hill and 5 = Lakes and rivers.

(Fig. 1). The local people have known of these exceptional landscape features for centuries. Sacrificial gifts have been brought to these lakes, and veneration has been shown by naming sacrificial stones (seitas) on their shores (Kotivuori & Torvinen 1993). The saivo lakes continue fascinate people up to present time, and these lakes have become popular attractions for geotourism.

Description and studies of the area

Multiple glacial advances and retreats have been described in the Äkäsaivo and Pakasaivo region (67^0 37 ‘ 00’’ N and 23^0 47’ 45’’ E) (Hirvas 1991). The Early Weichselian till bed and till-covered eskers provide evidence of an ice flow from northwest to southeast. Pakasaivo and Äkäsaivo are located on the northern edge of the ice divide that existed at the end of the Late Weichselian glaciation. The last flow direction of the continental ice sheet was towards the north-northeast. The region was deglaciated approximately 10 000 years ago (Johansson & Kujansuu 2005). In the vicinity of the saivos, there are glaciofluvial features typical of the supra-aquatic area, such as sharp-crested, steep-sided and winding esker ridges, alternating with gorges and channels, eroded deep into the glacial deposits, in places even into the bedrock. These features are evidence of the action of abundant meltwater generated by the melting of the continental ice. Pakasaivo, which is about one kilometre long and in places only tens of metres wide, is a canyon lake located in a bedrock fracture zone (surface 218 m a.s.l.). At the southern end, its shores are fairly gentle and stony. In the middle, the shores get steeper and the

Fig. 2. At the northern end of Pakasaivo, the steep rock walls rise up to 40 metres above the surface of the lake. (Photo: P. Johansson, 2000)

lake narrows between 20 – 25 metres high rock walls. At the northern end, the lake widens into a circular hollow, surrounded by steep rock walls and 120-140 metres in diameter. Due to the cubical fracturing of the granitic bedrock (Lehtonen 1981), the cliffs around Pakasaivo form step-like levels, separated by vertical walls (Fig. 2).

Äkäsaivo is a closed canyon lake (surface level 237 m a.s.l.), formed in a north-south fracture zone branching out from the Äkäsjoki river valley. The lake is bounded by steep, 30 metres high walls of till and bedrock (Fig. 3). Erosional remnants similar to tor formations rise on the shores, separate from the rock walls. The most special feature is the Seitakivi, a 20 metres high, pillar-like rock formation. It is known to have been a sacrificial site of the Sámi people (Kotivuori & Torvinen 1993).

In the summer of 2000, the Geological Survey of Finland carried out basic mapping of Quaternary deposits and related studies in the Yllästunturi region.

Fig. 3. Äkäsaivo with the surrounding steep rock walls in the Äkäsjoki river valley. (Photo: P. Johansson, 2000)

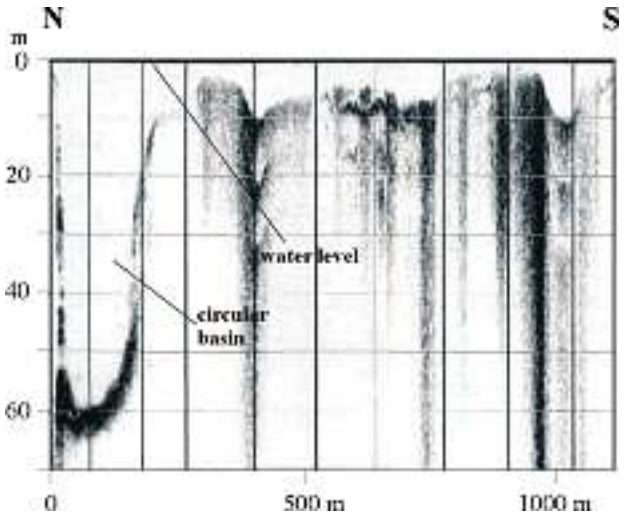


Fig. 4. Longitudinal profile obtained by echo sounding of the Pakasaivo canyon lake. The pothole-like hollow is 60 m deep.

The depth of Pakasaivo and its bottom topography were studied by echo sounding from a boat (equipment Furuno FE-881 MK II) (Johansson 2003). The sounding profiles show the water depth at the southern end and narrow middle of Pakasaivo to range, on average, between two and five metres and in the basin-like depressions between eight and ten metres. At the circular northern end of the lake, the bottom dives to a depth of tens of metres only a few metres from the shore, and in its middle the soundings show the depth to be 60 metres (Fig. 4). From all directions, the submerged rock walls slope steeper than 50 degrees towards the centre of the lake. The bottom is bedrock and a gyttja layer over seven metres thick and blocks that have tumbled down from the steep canyon walls. Preliminary observations indicated that the gyttja is composed of fragments of ligneous and herbaceous plants, such as leaves and pine needles (A. Ojala, personal communication 2005). The total depth of the canyon, including the 30-40 metres high rock walls on land, is over 100 metres, making it the second deepest lake in northern Finland (Karlsson 1986, Johansson 2003). Pakasaivo is classified as a meromictic lake, as it undergoes incomplete mixing of its water column during spring and autumn overturn periods (Wetzel 1983). The thermocline, or the layer between the epilimnion and hypolimnion, is at a depth of five to ten metres. Here an abrupt change in the oxygen content occurs, and the hypolimnion is oxygen-free and sulphurous. A study by the Lapland Regional Environment Centre in 1993 showed the oxygen content of the water to be 9 mg O_2/litre at a depth of five metres and only 0.8 mg at a depth of ten metres. Deeper down, the water was oxygen-free (P. Räinä, personal communication 2005).

Subglacial environment of formation

The formation of Pakasaivo and Äkäsaivo was a result of several geological processes. Fracture zones, formed between intact parts of the bedrock millions of years before the Ice Age, controlled the location and form of the saivos. As a result of glacial erosion over several glaciations, the bedrock around the fracture zones was eroded into depressions. During the last deglaciation, the meltwater from the inland ice formed a subglacial meltwater system, where the water flowed with great force in the direction of the pressure gradient towards the ice margin in the north. Pakasaivo is related to the glaciofluvial system that crossed the top of the nearby Karhulaki hill (370 m a.s.l.), creating glaciofluvial erosion and deposition features on the hillslopes (Johansson 2003).

Early during deglaciation, the glaciofluvial action was mainly erosion. The canyon of Keinokursu, which is incised into Karhulaki in the direction of the meltwater tunnel, was formed. It is over four kilometres long, 50 - 150 m wide and 30 m deep. The tunnel continued in the direction of the fracture zone across Pakasaivo (Fig. 5). At the northern end of Pakasaivo, a powerful subglacial meltwater whirl, evidently of long duration, was formed in the meltwater stream. The whirl ground a deep, pothole-like hollow into the broken rock. Although the walls show no distinct surfaces ground by the water whirl (cf. Shaw 1996), the form and exceptional depth of the saivo are clear evidence of the subglacial meltwater whirl. West of Pakasaivo, in northern Sweden, there are canyon-like valleys known as kursu valleys, major erosional features eroded into broken bedrock by enormous masses of meltwater. It is probable that the meltwater streams have also found their way into the same fracture zones in the bedrock during earlier deglaciations (Saarnisto 1991). Rudberg (1949) and Olvmo (1989) suggested that the kursu valleys were formed subaerially. In the cases of Pakasaivo and Keinokursu, the subaerial alternative is not possible since the subglacial esker ridges related to these canyons could only be formed at the base of the ice in a completely water-filled conduit, where the meltwater stream, subjected to high hydrostatic pressure, was in balance with the pressure caused by the surrounding ice sheet (Clark & Walder 1994, Brennand 2000). Evidence of subglacial origin is also evident in the fact that the meltwater stream did rise uphill (cf. Shreve 1972, Röthlisberger 1972, Nye 1973).

As the deglaciation proceeded, the location of the

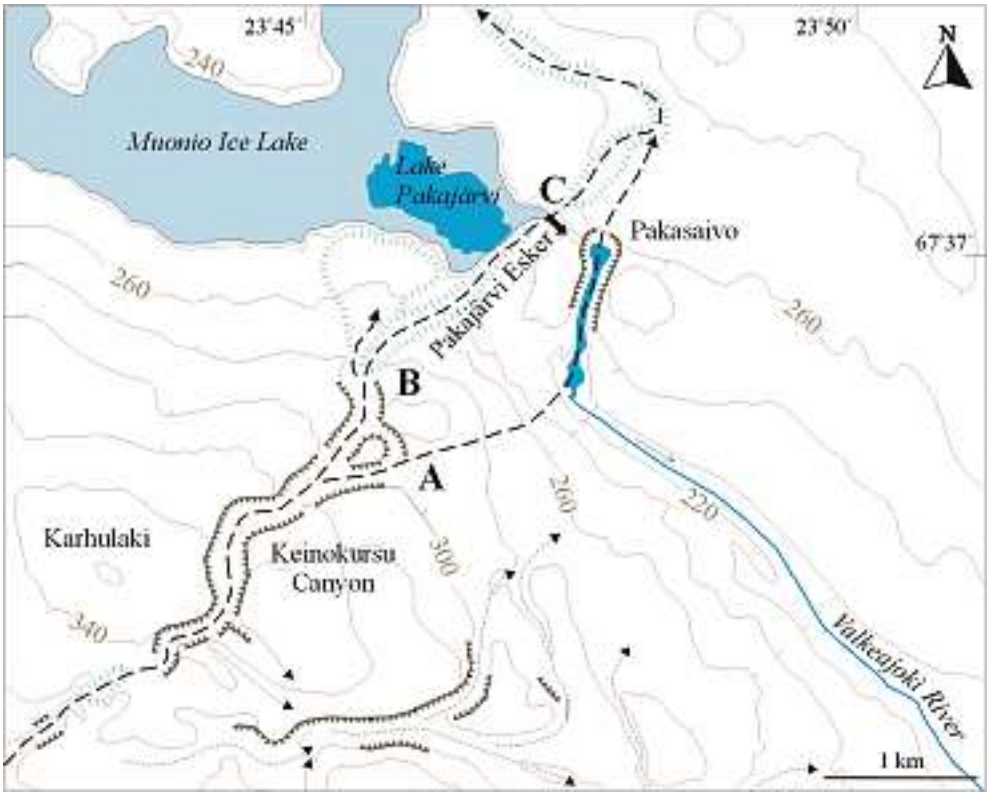


Fig. 5. Glaciohydrographic history of the Pakasaivo area. A = Route of the subglacial meltwater stream during its early stage, B = Route of the subglacial meltwater stream during its late stage, C = Spillway during the ice lake stage.

subglacial meltwater tunnel shifted to the north of Pakasaivo. This happened before the decrease in the meltwater pressure due to the thinning of the ice sheet caused the formation of an esker on the tunnel floor (cf. Lundqvist 1979, Ashley et al. 1991, Clayton et al. 1999). The Pakasaivo canyon must have been covered by the ice even earlier, since this is the only way the sediments transported by the glacial river could have been kept from accumulating there. As a result of the subglacial accumulation process, the almost 40 metres high Pakajärvi esker was deposited on the shore of the Pakajärvi lake about 500 metres from Pakasaivo (Fig. 5) and continued in the direction of Lake Kangosjärvi (Fig. 1). The Pakajärvi esker is part of a meltwater system from the last deglaciation that starts at the ice divide of central Lapland and continues north towards the ice margin (Johansson & Kujansuu 2005).

The features of Äkäsaivo and surrounding slopes also reflect powerful meltwater erosion. Most probably it was also subglacial based on rock walls that were worn by whirling water, and the pillar-like erosion remnants. In the Äkäsjoki river valley, north of Äkäsaivo, there are scattered glaciofluvial sand fields without a distinct morphology typical of eskers. They are not related to the other glaciofluvial formations from the last deglaciation in the area.

Erosion caused by the discharge from the ice lakes

During the last stage of deglaciation, meltwater was dammed in front of the retreating ice sheet and the Muonio Ice Lake was formed (Kujansuu 1967). At its largest the ice lake covered an area of about 800 km^2 on the Finnish side of the Muonio river valley. From the lake, spillways of various ages and at various altitudes led eastwards to the Ounasjoki river valley (Kujansuu 1967). After the retreat of the ice margin to the south-southwest from the Äkäsjoki river valley, the water of the ice lake pushed into the valley. For some time the Äkäsjoki river served as an ice lake spillway, and the meltwater flowed southwards causing erosion in Äkäsaivo. Glaciofluvial and till materials deposited in the valley were transported away, and the erosion continued down to the hard bedrock.

Due to the retreat of the ice margin, the ice lake spillway shifted westwards to the neck of land between Pakasaivo and Pakajärvi. In Pakasaivo, the erosive action of the water stream was more powerful than in Äkäsaivo. Water discharged from the ice-dammed lake first broke off the Pakajärvi esker, and then plunged into the hollow of Pakasaivo (Fig. 5). The rock wall bounding it in the north served as a threshold,

controlling the water level of the ice lake behind it at a level of 238 metres. As the water was discharged across the threshold, a 20 metre high cataract was formed. In fact, the form of the steep northern shore of Pakasaivo resembles a cataract lip and the deep hollow of the lake itself may be a plunge pool lake formed at the base of the cataract (Johansson 2003). The water rushing into the hollow possibly yielded vertical vortices that caused a turbulent plucking action. Similar deep hollows created by the discharge of ice-dammed lakes are also found for example in the Channeled Scabland in the state of Washington, USA (Baker 1978, 1987).

Conclusion

With regards to their formation, the saivo lakes of Pakasaivo and Äkäsaivo in western Finnish Lapland are polygenetic. They are located in major fracture zones of the bedrock, where the rock was deeply fractured between intact blocks since preglacial times. The direction of the fracture zones had a clear influence on the orientation of the saivos. In the formation of Pakasaivo and Äkäsaivo, glacial erosion also played an important role, but the most important erosional factor was the pressurized meltwater stream flowing in the subglacial conduit and the powerful subglacial whirling it caused. The location of Pakasaivo is also caused by the route of a subglacial meltwater system, stemming from the last deglaciation and running from south to north. Near Äkäsaivo, no traces of a meltwater system of the same age are found. The nearest eskers, at Nivunkijärvi and in the Lainiojoki valley, belong to a meltwater system in the northwest-southeast direction that is older than the last deglaciation. Along with the eskers, the sand fields in the Äkäsjoki river valley and Äkäsaivo may be related to the same system. At least they are situated in the same northwest-southeast line, which is parallel to the older direction of glacier flow (Hirvas 1991). This line crosses the Ylläs-Lainiotunturi mountain range. In the mountain area, no eskers or other signs of glaciofluvial action are found, which may be due to the strong relief of the terrain. Erosional features formed by a meltwater stream may exist here, but they have been covered under soil deposits accumulated by the last glaciation. Äkäsaivo is an exception, as the canyon is too deep to be covered. During the last stage of the last glaciation, the water discharged from the Muonio Ice Lake only finished the work of the subglacial meltwater erosion. In Pakasaivo, it had a considerable effect on the shape of the hollow and the steep cliff on its northern shore. The powerful stream cleaned the slopes and bottom of the saivos from accumulated rock fragments.

REFERENCES

Ashley, G.M., Boothroyd, J.C. & Borns, H.W. Jr. 1991. Sedimentology of Late Pleistocene (Laurentide) deglacial-phase deposits, eastern Maine: An example of a temperate marine grounded ice-sheet margin. In: Anderson, J.B. & Ashley, G.M. (eds.), Glacial marine sedimentation: Paleoclimatic significance. Geological Society of America, Special Paper 261, 107-125.

Baker, V.R. 1978. Paleohydraulics and hydrodynamics of scabland floods. In: Baker, V.R. & Nummedal, D. (eds.), The Channeled Scabland: A guide to the Geomorphology of the Columbia Basin, Washington. Planetary Geology Program, National Aeronautics and Space Administration, 59-79.

Baker, V.R. 1987. Dry falls of the Channeled Scabland, Washington. Geological Society of America Centennial Field Guide – Cordillieran Section, 369-372.

Brennand, T.A. 2000. Deglacial meltwater drainage and glaciodynamics: inferences from Laurentide eskers, Canada. Geomorphology 32, 263-293.

Clark, P.U. & Walder, J.S. 1994. Subglacial drainage, eskers, and deforming beds beneath the Laurentide and Eurasian ice sheets. Geological Society of America, Bulletin 106, 304-314.

Clayton, L., Attig, J.W. & Mickelson, D.M. 1999. Tunnel channels formed in Wisconsin during the last glaciation. In: Mickelson, D.M. & Attig, J.W. (eds.), Glacial processes Past and Present. Geological Society of America, Special Paper 337, 69-82.

Hirvas, H. 1991. Pleistocene stratigraphy of Finnish Lapland. Geological Survey of Finland, Bulletin 354, 123 p.

Johansson, P. 2003. Eskers and bedrock gorges (tunnel valleys) in the Pakasaivo area, western Finnish Lapland. Bulletin of the Geological Society of Finland 75, 5-15.

Johansson, P. & Kujansuu, R. (eds.) 2005. Pohjois-Suomen maaperä: maaperäkarttojen 1:400 000 selitys (Quaternary deposits of Northern Finland – Explanation to the maps of Quaternary deposits 1:400 000). Espoo: Geologian tutkimuskeskus, 236 p. [in Finnish]

Karlsson, K.-P. (ed.) 1986. Atlas of Finland, folio 132, Water. National Board of Survey and Geographical Society of Finland, Helsinki, 31 p.

Kotivuori, H. & Torvinen, M. 1993. Tunturi-Lapin kiinteät muinaisjäännökset. Lapin Seutukaavaliitto A 130, 119 p.

Kujansuu, R. 1967. On the deglaciation of western Finnish Lapland. Bulletin de la Commission Géologique de Finlande 232, 96 p.

Lehtonen, M. 1981. Kihlanki. Geological Map of Finland 1:100 000, Pre-Quaternary Rocks, Sheet 2714. Espoo: Geological Survey of Finland.

Lundqvist, J. 1979. Morphogenetic classification of glaciofluvial deposits. Sveriges Geologiska Undersökning C 767, 71 p.

Nye, J.F. 1973. Water at the bed of a glacier. Symposium on the Hydrology of Glaciers, Cambridge, 9-13 September 1969. Publication de l'Assosiation Internationale d'Hydrologie Scientifique 95, 189-194.

Olvmo, M. 1989. Meltwater canyons in Sweden. A study of canyons of the "kursu", "skura" and "grav"-type. University of Göteborg, Department of Physical Geography, Rapport 27, 126 p.

Paulaharju, S. 1922. Lapin muisteluksia. Porvoo: WSOY, 231 p.

Röthlisberger, H. 1972. Water pressure in intra- and supraglacial channels. Journal of Glaciology 11, 177-203.

Rudberg, S. 1949. Kursudalar i Norrbotten, en preliminär översikt. Meddelande Uppsala Universitetets Geografiska Institutionen A 64, 442-495.

Saarnisto, M. 1991. Tornionlaakson geologiset kehitysvaiheet. In: Hederyd, O., Alamäki, Y. & Kenttä, M. (eds.), Tornionlaakson historia I. Jääkaudelta 1600-luvulle. Mahlung: Malungs Boktryckeri, 11-44. [in Finnish]

Shaw, J. 1996. A meltwater model for Laurentide subglacial landscapes. In: McCann, S.B. & Ford, D.C. (eds.), Geomorphology Sans Frontières. Chichester: Wiley & Sons, 181-236.

Shreve, R. 1972. Movement of water in glaciers. Journal of Glaciology 11, 205-214.

Wetzel, R.G. 1983. Limnology. Philadelphia: Saunders College Publishing, 767 p.

Quaternary studies in the northern and Arctic regions of Finland
Edited by Antti E.K. Ojala
Geological Survey of Finland, Special Paper 40, 41– 51, 2005.

GLACIAL HYDROLOGY AND THE DEGLACIATION OF THE SCANDINAVIAN ICE SHEET IN THE PAISTUNTURIT AREA OF THE UTSJOKI REGION IN NORTHERN FINNISH LAPLAND

by
Riikka Koskinen

Geological Survey of Finland, P.O. Box 96, FIN-02151, Espoo , FINLAND
E-mail: riikka.koskinen@gtk.fi

Koskinen, Riikka, 2005. Glacial hydrology and the deglaciation of the Scandinavian Ice Sheet in the Paistunturit area of the Utsjoki region in Northern Finnish Lapland. *Geological Survey of Finland, Special Paper 40*, 41–51, 5 figures and 1 table.

A glacial geomorphological investigation into the deglaciation of the Paistunturit fell area in Northern Finnish Lapland is synthesised in this article. Glacial meltwater drainage characteristics were studied through aerial photograph interpretation and the results were utilized in mapping various stages of deglaciation of the Scandinavian ice sheet in the Paistunturit area. Landforms created by glacial meltwater erosion and deposition were identified and classified and the results were applied in describing the paleohydrography of the melting ice sheet in the study area. Because the study area is supra-aquatic and periglacial processes and wind erosion have had little or virtually no effect on the landforms since deglaciation occurred, the glacial landforms are especially well preserved. The vegetation cover is sparse and therefore the landforms can be easily observed from aerial photographs. The ice sheet margin receded towards the south and southwest in the study area. The gradient of the retreating margin of the ice sheet was determined to be 3/100 and the annual downwastage of the glacier surface 1.7–3 m. The glacier snout retreated 60–100 m yv^{-1}. On the basis of the vertical and horizontal components, the deglaciation of the study area was estimated to have lasted between 200–300 years.

Key words (GeoRef Thesaurus, AGI): glacial geology, Scandinavian ice sheet, deglaciation, glacial features, paleohydrology, meltwater channels, lateral channels, Paistunturit, Utsjoki, Finland

Introduction

The identification, classification and description of glacially-derived erosional and depositional landforms have great significance in regional deglaciation studies. The pattern of the former flow of glacial meltwaters can be reconstructed by studying the morphology of the glacial meltwater channels (Kujansuu 1967, Sugden et al. 1991, Johansson 1995, Glasser et al. 2004). As stated by Kujansuu (1967), the mode of deglaciation and the glaciological state of the ice sheet are best reflected, in addition to lateral drainage channels, by the evidence of both depositional and erosional subglacial activity. Parts of Finnish Lapland as well as the Paistunturit fell range have remained supra-aquatic and the glacial landforms have undergone little reworking through periglacial processes and wind erosion since deglaciation (Kujansuu 1992, Eronen & Haila

1990). Because the vegetation cover is sparse, aerial photograph interpretation is relatively easy.

Although the deglaciation of Finnish Lapland has been previously studied in great detail, little research concerning the the deglaciation of the northernmost areas of Finland and especially the Paistunturit area of the Utsjoki region has been done. This is partly due to the remote location of the area. The geology and geomorphology of the northernmost Finnish Lapland, the Utsjoki and the Inari region, have been studied by Tanner (1915), Mansikkaniemi (1967, 1970, 1973), Hyvärinen & Eronen (1975), Heikkinen & Tikkanen (1979), Seppälä (1971, 1980), Porola (1981), Honkala (1982) and Riipinen (1994). The glacial history of the Paistunturit fell range has been investigated by Syrilä (1964, 1970), Karzcewski (1975) and Kaitanen (1969, 1989). Nevertheless, the deglaciation history of the Paistunturit fell range has not previously been studied to the same extent as presented by Koskinen (2000). This article synthesises the results obtained in the study.

A great abundance of melt water channels, such as overflow channels, exists in the study area. Because the study area is located directly south of the Younger Dryas ice-marginal formations, the Tromso-Lyngen formation, which was formed 11 680–10 150 BP (Erikstad 1998), the Paistunturit area has great significance in the deglaciation history of the northernmost areas of Finnish Lapland.

Study area and methods

The investigated Paistunturit fell area is located in the Precambrian granulite fell zone of the Utsjoki region of northernmost Finnish Lapland. The granulite complex consists of separate blocks divided by

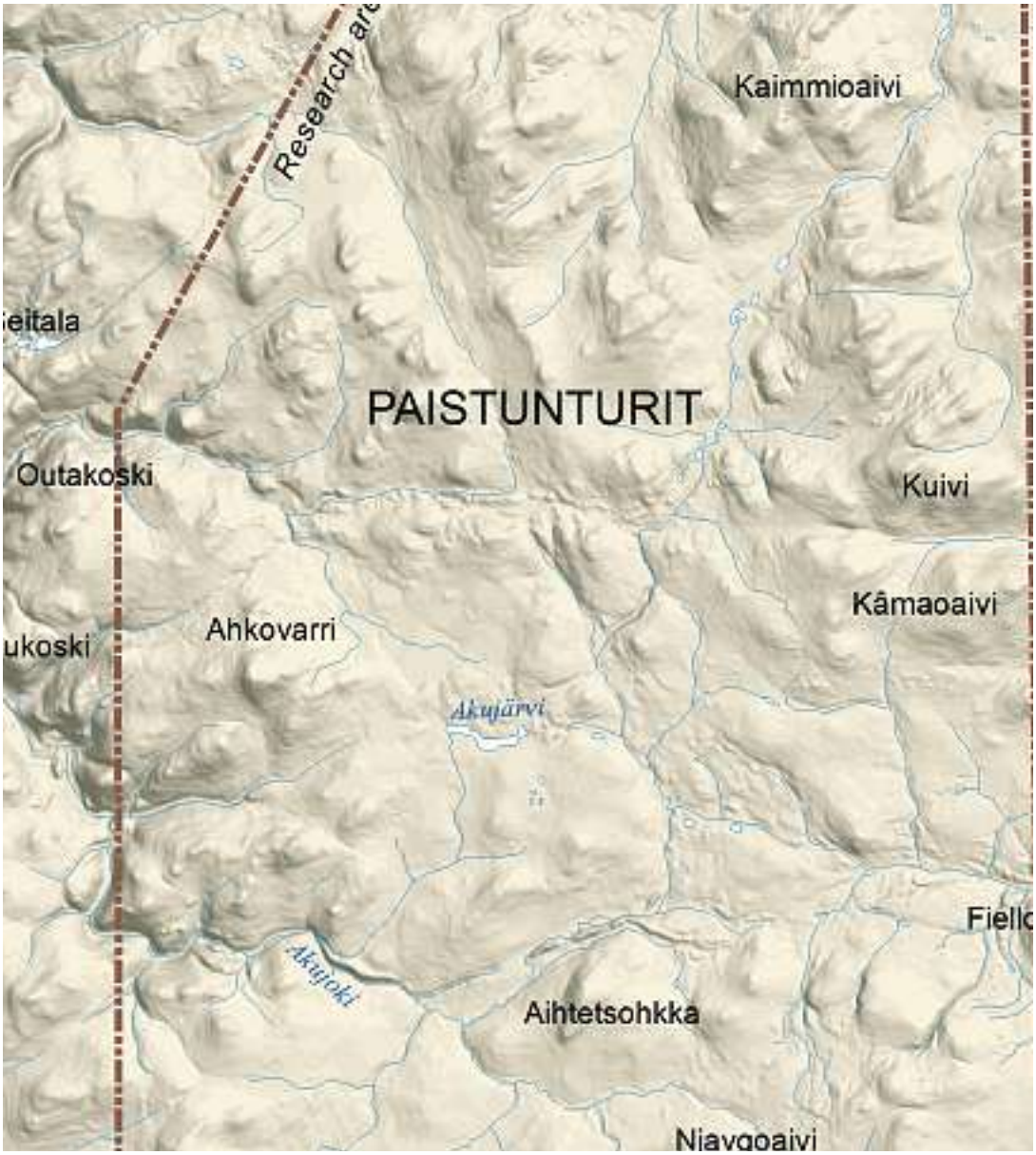


Fig. 1. The location of the reseach area

bedrock fracture zones (Tanner 1938, Meriläinen 1976). The gently sloping mountain area, referred to as "*the upland peneplain"* (Mikkola 1932), is bound to the west by the Teno River valley and to the east by the Kevo Strict Nature Reserve. Only a few summits with altitudes exceeding 600 m exist in the area and the general height is 300–400 m above sea level (Kaitanen 1989). The southern boundary of the study area is 69º 30' 41" N, the northern boundary 69º 42' 30 N, the eastern boundary 26º 29' E and the western boundary 26º 03' E (Fig. 1).

Mapping of glaciated terrain traditionally consists of the identification of landforms according to origin and age (Benn & Evans 1998, Evans 2003). The main method used in the study was aerial photograph interpretation. Surface landforms created by meltwater erosion and deposition were identified and classified from 45 black and white aerial photographs of 1:30 000 and 1:15 000 in scale. The identification and classification of the glacial meltwater channels was carried out according to their size, shape, gradient, location and position with respect to the ice sheet, and its margin, relationship with depositional formations and the visible tone or shading of the landforms on the aerial photographs. Other criteria used in identifying the meltwater channels included underfit drainage, an abrupt initiation and termination, a hanging position, a convex or undulating long profile, and the crossing of water divides, i.e. erosion and deposition had occurred on a upward sloping plane (Price 1973).

The meltwater channels were classified on the basis of distinctive channel morphologies into lateral drainage channels, submarginal channels, overflow channels, subglacial channels and extramarginal channels. The identified features were transferred onto a map at a scale of 1:50 000. Fieldwork carried out during two field seasons in the late summers of 1998 and 1999 involved verifying the aerial photograph interpretation and obtaining information on the size, shape, location, elevation and composition of the studied formations.

Special emphasis was placed on carrying out level and gradient measurements on the lateral drainage channels from the aerial images to determine the direction of the flow of the meltwaters at the glacier margin, the surface gradient of the retreating ice margin and the rate of deglaciation. The measurements enabled determination of the annual rate of retreat of the glacier snout in the mountain area. The final deductions of the terminal positions of the retreating glacier margin were carried out on the basis of observations and measurements gathered from the erosional and depositional landforms, and especially the lateral meltwater channels and overflow channels. Evidence of ice dammed lakes was utilized in the delineation of the various stages of retreat of the glacier margin. Subglacial paleohydrography was studied by examining the landforms attributed to subglacial meltwater flow.

Results

Deglaciation of the study area began when the summits of the mountain range began to emerge as nunataks above the surrounding ice sheet in the northern and north-eastern corner of the study area. The highest peaks, Kuivi (640.5 m) and Kaimmioaivi (618.9 m), were the first summits that emerged from under the ice. According to Kaitanen (1969), the surface of the ice sheet was only 100–200 m above the highest summits before the retreat of the ice margin began from the Younger Dryas End Moraines. Kaitanen (1969) also suggested that during the last glacial maximum the ice sheet was only slightly thicker and glacial erosion was weak. Abundant evidence in support of Kaitanen's findings of the weakness of glacial erosion exists in the area in the form of bedrock residuals, so called "Tors", in the thickness of the weathered bedrock and in the general lack of landforms generated by glacial erosion, such as drumlins and flutings typical of Inari basin southwest of the Paistunturit fell range (see e.g. Mikkola 1932, Kaitainen 1969, 1989, Karczewski 1975).

Proglacial drainage

The ice masses melted slowly at the beginning of the deglaciation because of the prevailing cold climatic conditions and reduced precipitation levels. The reduced precipitation levels resulted from the reduction in the moisture-bearing capacity of the surrounding atmosphere due to the lower temperatures. Likewise, the aerial extent of the nunataks emerging above the glacier surface was small and therefore the amount of solar radiation absorbed by the darker nunatak surfaces was low and its heating of the surroundings had little or no effect on the ice melt rates. The peaks of the summits were eroded by fluvial erosion, periglacial processes and wind and glacial erosion. Melting was restricted to the surface of the glacier and meltwater flow to marginal and englacial drainage routes (see e.g. Johansson 1990, 1994, 1995).

Judging from the abundance of drainage channels, meltwater had been present at the glacier margin and underneath it since the early stages of deglaciation. Therefore, the glacier ice had been above the pressure melting point and the quantities of meltwater had been sufficient to produce channelized drainage. Although

the climate was cold and a moisture deficit existed during the early stages of deglaciation, the conditions did not resemble those of modern polar continental ice margins where sublimation exceeds precipitation and where ablation mainly consists of sublimation (see e.g. Fitzimons 2003).

As the meltwaters found their way through passes between the peaks of the highest summits, overflow channels were formed. The overflow channels show evidence of severe fluvial erosion with the channels typically incised through the surficial deposits to the bedrock. Steep-walled and relatively short channels are often located in preglacial bedrock fracture zones (see e.g. Penttilä 1963, Kujansuu 1967, Syrilä 1970, Abrahamsson 1974, Johansson 1990, Kujansuu 1992). Most of the overflow channels are located at altitudes between 400–500 m; they typically open up in the direction of the inclination of the ice sheet margin, i.e. towards the north-east and north, with some exceptions that can be attributed to differences in the direction of retreat of the glacier margin caused by the bedrock relief and differences caused by bedrock fracture zones that guided the flow of the meltwaters.

The Kuivi ice lake

A small marginal ice-dammed lake was formed between the glacier margin and arch-like ridges of the Kuivi and Guovdoaivi fell during the nunatak phase. Evidence of the mainly supraglacial body of water can be found in the form of deep spillways leading east, north-east and north through passes at the summit of Kuivi and south-western flanks of Guovdoaivi fell. The overflow channels were formed as the meltwaters, that had collected behind the nunataks burst through the passes between the summits. Because the upper slopes were exposed to heavy erosion during the early stages of deglaciation, evidence of shore marks has been destroyed. Another explanation for the lack of shore marks is the short life-span of the water bodies. Therefore, shore marks were not developed at all.

As the Kuivi icelake was most likely partly supraglacial, it is hard to estimate its total areal extent. The oldest of the overflow channels, i.e. the first route of the meltwater flow through the summit passes, is located in a bedrock fracture zone on the summit of Kuivi Fell at the altitude of 560 m (Fig. 2). The eastward flow of the meltwaters became partly subglacial on the eastern flank of the Kuivi fell as the meltwaters flowed under the glacier margin eroding a deep subglacial chute. The water level in the lake gradually sank as the ice sheet margin thinned and the eventually new outlet channels were eroded at the altitudes of 500, 490, 480, 470, 455, 450, 440, 337.5 and 335 m.

Lateral drainage

Apart from the overflow channels, meltwater channels are rare in the summit areas which, as mentioned earlier, were exposed to heavy erosion during the nunatak phase of the deglaciation. Simultaneously with the recession of the ice margin, the surface of the ice sheet successively sank. The meltwaters flowed at the edge of the glacier margin at the beginning of the

Fig. 2. The oldest of the overflow channels, i.e. the first route of the meltwater flow through the summit passes, is located in a bedrock fracture zone on the summit of Kuivi Fell at the altitude of 560 m a.s.l. (Photo: R. Koskinen, 1998)

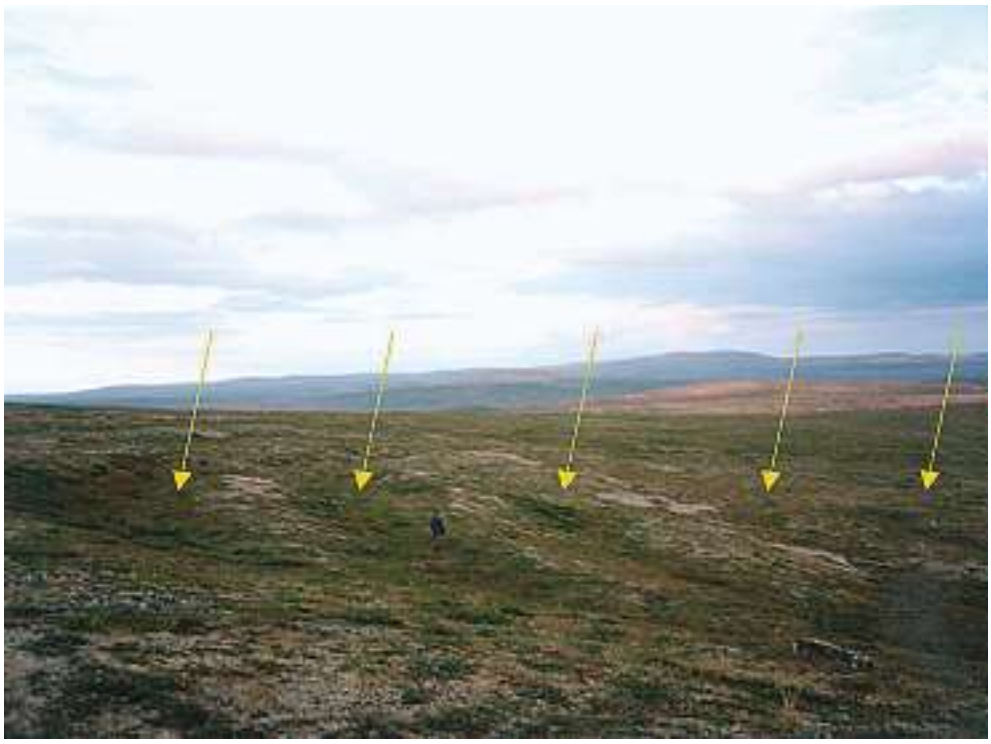


Fig. 3. Lateral meltwater channels on the southern flank of Akuvaara fell. (Photo: R. Koskinen, 1998)

melt season, when frozenbed conditions existed as a result of the cold wave penetrating the thin ice-margin during the winter accumulation season. Cold-based ice, however, occurred only at the very margin of the glacier. Because of the presence of permafrost during the deglaciation phase, it is likely that the meltwaters also excavated routes in the ice itself rather than in the frozen bedrock or the frozen till overlying the bedrock. Later during the melt season, the meltwater flow became submarginal. Gradually, the subglacial channels, or chutes, opened up and the melt waters began to flow underneath the margin and towards the bottom of the glacier following the steepest hydraulic gradient. The chutes run at right angles across the lateral drainage channels. Regarding the gradient of the glacier margin and the vertical intervals of ice-marginal lateral channels on the upper parts of the summits, having an average value of 1.7 m, the glacier margin retreated 60 m per year towards the south and south-west at the beginning of the deglaciation phase. Gradually, more and more nunataks emerged above the surrounding ice sheet.

As the deglaciation progressed, the glacier margin became more undulating due to the influence of the local relief on the mode of retreat of the glacier margin. A series of gently downward sloping lateral drainage channels was formed as the meltwaters flowed between the mountain sides and the glacier margin (Fig. 3). The lateral channels were formed during the spring and summer melt seasons and the meltwater routes alternated between marginal and submarginal; thus simultaneous erosion of marginal and submarginal channels has been possible. It is also possible that meltwaters used the same routes during more than one melt season. Although the lateral channels reflect the gradient of the glacier margin and the rate of downwastage, the proposal of their annuality (e.g. Mannerfelt 1945, 1949, Sissons 1958) should therefore be treated with caution.

The large series of lateral channels typical of central Lapland is rare in the study area as a result of the less favourable topographic conditions in relation to the direction of retreat of the glacier margin. The best developed series of channels in the Paistunturit area are generally located at altitudes between 350–450 m on southern and south-eastern slopes having gentle gradients, on mountain sides facing the glacier margin and overlaid by relatively thick till cover (Table 1). Kaitanen (1969) made similar observations of the complete lack of lateral drainage channels above the altitude of 500 m. Drainage channels are rare on north-eastern and eastern slopes, on the flanks receiving less insolation. Meltwater channels are found on the eastern flanks only in valleys that open up towards the north, except for the channels located on the eastern flank of the Kuivi fell. The length of the channels varies between 300–1200 m and they begin and terminate abruptly. The drainage channels vary

Fig. 4. A series of lateral and submarginal meltwater channels on the southern flank of Akuvaara fell at the altitude between 350-460 m. The series consists of approximately 31 channels that can best be classified as being strictly lateral at altitudes between 420-460 m. The vertical intervals of the lateral meltwater channels on the upper part of the flank are 1.7 m. The average interval between the channels on the lower part of the flank is 2.5 m. The average gradient of the channels is 3.2/100 (Copyright Topografikunta, permit n:o 55/2005).

in size, shape and depth depending on their altitude. On the upper slopes the lateral channels are usually shallow, one- or two-sided, their shape is irregular and, because of destruction caused by solifluction, the channels are difficult to identify and classify from aerial photographs (see e.g. Syrilä 1964, 1970). On slopes with steep gradients the channels resemble stony terraces.

The lateral drainage channels become larger and their vertical intervals grow on the lower parts of the flanks. The growth of the vertical intervals resulted from the influx of meltwater leading to increased erosion rates as the ice melt rates increased during the deglaciation phase. Also, as the ice surface remains steepest in its marginal parts and the gradient decreases gradually towards the center of the ice sheet, referred to as the ice divide, the gradient of the surface was less at the time the first nunataks emerged above the ice sheet. Therefore, the gradient of the lateral channels on the upper slopes is smaller than on the lower flanks. The erosional energy of the meltwaters flowing off the surface of the ice sheet margin or along side it was also weaker during the nunatak phase than later on during the deglaciation because of the lesser gradient of the margin.

As the channel gradients become larger, the lateral channels grade into submarginal and eventually subglacial channels (Fig. 4). It is common that the lateral, submarginal and subglacial channels form a web-like network on the lower flanks and reliable identification and classification of the channels therefore becomes difficult and sometimes impossible (see e.g. Hyvärinen & Eronen 1975). According to Kujansuu (1967), this is typical for channel series formed on mountain sides facing the glacier margin, as for the channels series in the study area. Active solifluction processes complicate channel identification and classification.

The problems of the eskers

Regarding the gradient of the glacier margin of 3/100 and the vertical intervals of ice marginal lateral channels on the lower flanks, having an average value of 3 m, the glacier margin retreated approximately 100 m per annum as the retreating glacier margin reached the central and southern parts of the study area. The gradient has been sufficient to maintain slow movement of the ice masses towards the margin, thus preventing the ice masses from becoming semi-stagnant or stagnant. Although the glaciological state of the ice sheet was deteriorating, a sharp-crested esker, the Sidnokoveaja esker, was deposited in the central part of the study area. The segmented esker was deposited as subglacial tunnel fill and therefore recorded the route

Table 1. The average vertical intervals and gradients of the lateral drainage channels in Finnish Lapland (Abrahamsson 1974, modified by the author).

Area	Location	General Altitude	Author	Average Vertical intervals	Average Gradient
Sodankylä	67° N 27° E	190–220	Tanner (1915: 647)	1,5 m	2–4/100
Sodankylä	67° N 27° E	180–200	Tanner (1915: 647)	0,7–1 m	2–4/100
Sodankylä	67° N 27° E	n. 200	Tanner (1915: 647)	0,52 m	2–4/100
Sodankylä	67° N 27° E	220–240	Tanner (1915: 647)	0,8 m	2–4/100
Yllästunturi	67° 30' N 24° E	500–600	Tanner (1915: 647)	1–2 m	2–4/100
Tankavaara	68° N 28° E	350–450	Virkkala (1955: 411)	2,5 m	2/100
Laanila	68° 30' N 27° E	400–500	Penttilä (1963: 37)	1,6 m	1–4/100
Utsjoki	69° 30' N 26° E	380–440	Syrilä (1964: 340–341)	2 m	2–3/100
Marastotunturi	69° N 26° E	350–470	Piirola (1967: 31)	2 m	4/100
Enontekiö	69° N 22° E	–	Kujansuu (1967: 68, 87)	n. 2 m	1,7–2,7/100
Lainiotunturi	67° 30' N 24° E	500–600	Kujansuu (1967: 68, 86, 87)	2 m	2/100
Yllästunturi	67 30' N 24° E	350–450	Kujansuu (1967: 68, ,86)	3–3,7 m	2,2/100
Pyhätunturi	67° 30' N 24 °E	380–460	Kujansuu (1967: 68, 87)	3 m	–
Aakenustunturi	67° 30' N 24° E	–	Kujansuu (1967: 68, 87)	1,9–3,2 m	–
Utsjoki/ Paistunturit	69° 30' N 26° E	380–460	Syrilä (1964: 340, 341–342)	2 m	2–3/100
		350–450	Syrilä (1970: 22, 26),	0,5–2,5 m	1–5/100
		350–450	Koskinen (2000: 68–69, 88, 92)	1,7–3,0 m	3/100
Lainiotunturi	67° 30' N 24° E	380–590	Abrahamsson (1974: 159, 167)	2,4 m	3,7/100
Kukastunturi	67° 30' N 24° E	380–400	Abrahamsson (1974: 159)	2–3 m	–
Äkäslompolo	67° 30' N 24° E	380–590	Abrahamsson (1974: 159, 168)	1,7 m	4/100
Saariselkä	68° N 28° E	150–600	Piirola (1982: 268, 276)	1,63 m	4/100
Saariselkä	68° N 28° E	–	Johansson (1988: 550)	1–2,5 m	–
Saariselkä	68° N 28° E	280–480	Johansson (1995: 27–29)	1,2–3,5 m	1,5–3/100
Vuotso	68° N 27° E	–	Kujansuu & Hyyppä (1995: 48–49, 61)	1,53 m	2/100

of subglacial meltwater flow at the glacier bed (Benn & Evans 1998). As the ice movement rates gradually decreased, the subglacial tunnels might have been able to remain stable for years. Because the deposition occurred uphill, the esker comprises both erosional and depositional sections and therefore has an up-and-down grading long profile. The direction of the water flow in the conduit was from the west towards the east or north-east. The direction of the water flow and the hydraulic gradient in the pressurized conduits was controlled by the ice surface gradient and to a great deal by the form of the bed. The esker path, the subglacial tunnel in which the esker was deposited, follows the routes perpendicular to the equipotential surfaces on the glacier bed (Shreve 1985, Benn & Evans 1998). Deposition most likely took place close to the marginal zone of the glacier. Clear evidence of temporal changes associated with the formation of the esker exists in the form of sections comprised of Nye channels, i.e. channels incised into the substratum followed by depositional sections. Deposition probably occurred because of a temporal fall in discharge in the channel and on ascending planes as the pressure in the conduit fell. Erosion of the substratum resulted from the uphill flow of the meltwaters. According to Benn & Evans (1998), the presence of channels presented by Nye (1973) implies that the subglacial conduit was a relatively stable feature of the glacier bed as the water flow was consistently focused on the same route. This occurs in places where bedrock topography has a great role in defining the route of discharge, as is the case in the study area. Also, as stated by Shreve

(1985), the gentler the ice sheet surface gradient, as is the case in the study area, the more the esker path will be influenced by topography.

Another esker, the Akuharku esker, stretches across the southern part of the study area. Syrilä (1970) suggested that during the time the Akuharju esker was deposited the ice sheet would have been stagnant throughout its marginal parts. Because the both existing esker chains were deposited on an upward-grading plane it is unlikely that the glacier would have become stagnant in its marginal parts (glaciologically dead) at the time the eskers were formed. The waters flowed subglacially under hydrostatic pressure in pressurized conduits. If the ice sheet had been dead, such conditions could not have existed.

In the light of the results gained in this investigation, it is most likely that the Sidnokoveaja esker was deposited during the last deglaciation phase. However, the existence of a till cover on the top of Akuharju esker (see e.g. Syrilä 1970) suggests that it was deposited before the last deglaciation phase (see e.g. Johansson & Kujansuu 2005). Abundant observations of eskers having a till cover have been made in southern and central Fnnish Lapland (see e.g. Kujansuu 1967, 1975, 1994, Johansson & Kujansuu 1995, Johansson 2005). The remote location of the Paistunturit area has so far hindered research into the age, origin and structure of the esker systems.

The Akujärvi ice lake

As the undulating glacier margin reached the central parts of the study area (Stuorra-Mavdna-Tuoddar-Mavdna-Răsoaivi-Kămajokkeädšeoaivi-Kamaoaivi-zone), the northern and northeastern parts of the study area became free of the ice cover. Ice-dammed lakes were rare in the northern and north-eastern part of the study area resulting from the flow of the meltwaters away from the glacier margin following the downward sloping gradient of the relief in the valleys. In the central parts of the study area, small bodies of water gathered at the glacier margin forming ice-dammed lakes, because the glacier margin receded from higher to lower elevations. The marginal ice-dammed lakes were relatively small and shallow bodies of water with relatively poorly developed shore marks that can be at best followed for rather short distances (see e.g. Penttilä 1963, Kujansuu 1967, 1992, Seppälä 1980, Johansson 1988, 1990, 1995, Kujansuu & Hyyppä 1995, Johansson et al. 1997). This is due to the short life span of the lakes. The best-developed shore marks were formed relatively close to the glacier margin (see e.g. Lundqvist 1969, 1972, Johansson 1995). The shoremarks can be described as washing lines or surfaces and are generally quite hard to detect both from aerial photographs and in the field. In some cases, the slopes that lie under the shore lines are visibly darker in colour because the vegetation cover is denser. This is due to the deposition of finer lake-bottom sediments that are more able to withhold water.

In some cases the water bodies could be regarded as merely enlarged outlet channels or, as described by Kujansuu (1967), lateral bodies of water dammed up along the edge of the glacier margin. The existence of the lakes is best recorded in the presence of small delta-formations deposited at the end of spillways leading to the lakes. The largest of the ice lakes was dammed between the south-west retreating glacier margin and the Juohko-oaivi and the Ahkojavrkielas Fell in the low-lying Tšulloveijeäggi area. The surface altitudes of the ice-lake can be reconstructed on the basis of spillways crossing the central and north-eastern part of Juohko-oaivi fell at the altitudes of 402.5 m and 395 m. The spillways crossing Juohko-oaivi are quite shallow and therefore the meltwaters have flowed through them for only a short time period. As the retreat of the glacier margin proceeded towards the south-west, the surface of the ice lake sank. New spillways opened up on the south-western and southern side of the Ahkojavrkielas fell at the altitude of 375 m. While the glacier margin lay on the southern and south-western side of the Akuvaara fell, the ice lake filled the whole Tšulloveijeäggi basin, forming two lake basins that were connected by a narrow channel located on the eastern side of Luovdijeäggi fell (Fig. 5). New outlet channels opened up at the altitude of 360 m on the eastern side of the Akujärvi lake through which the meltwaters flowed towards the east. The deep and steep-walled channels show evidence of severe fluvial erosion.

Shore marks were formed on the western and north-western flanks of Eärnavâdda fell at the altitudes of 360–345 m and on the northern side of Akuvaara lake at the altitude of 360 m. During the final stages of existence of the ice lake its surface level dropped by 5 m to 355 m.

Small bodies of ice were cut off from the ice sheet margin during the deglaciation phase. Large marginal lobes typical of western Lapland (see e.g. Kujansuu 1967, Johansson 1990, 1995, Kujansuu & Hyyppä 1995) were not formed because of constrains caused by the relief. The valleys that open up in the direction of the retreat of the margin are relatively short in the study area. Small separate ice lobes lay on valley floors of the Mavnna river valley and the Mavnna-avdši-river valley in the northern part of the study area, leaving behind typical dead ice features in the

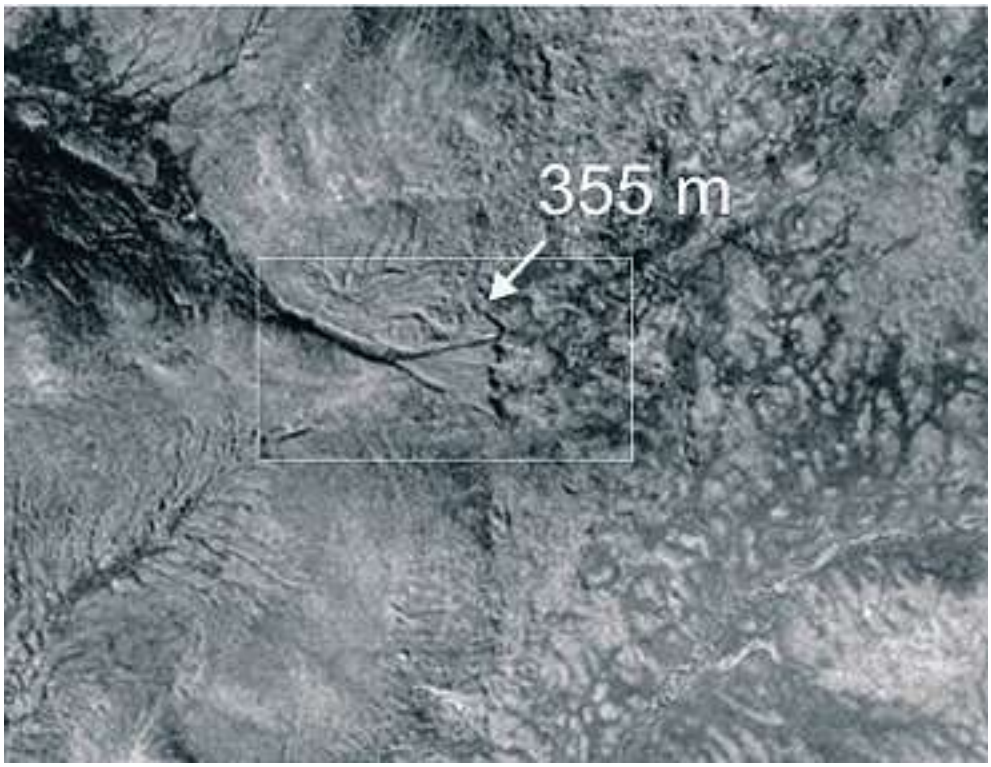


Fig. 5. A small delta was deposited in an ice-dammed lake south east of the Akuvaara fell. The delta is approximately 400 m long and 250 m wide and its surface altitude is 355 m a.s.l. (Copyright Topografikunta, permit n:o 55/2005).

form of hummocky moraines as the lobes gradually melted. Although small bodies of ice were separated from the margin of the ice sheet, the margin remained generally uniform throughout the deglaciation phase. During the final stages of deglaciation the valleys of the southern and south-western part of the study area were deglaciated.

Final remarks

Because there are too few well-developed series of lateral channels in the study area, it is difficult to estimate the exact mode and rate of retreat of the glacier margin in the mountain area. At present it is not exactly known when the deglaciation of the Paistunturit area took place. In view of the absence of research into the timing of the deglaciation, estimations regarding the event must be made on the basis of studies conducted in the near vicinity of the study area. According to Johansson (1990) the glacier margin retreated to central and western Lapland during the following 1200–1500 years after the Younger Dryas Stadial. Johansson's findings are consistent with estimation presented by Kujansuu et al. (1998), according to which the Inarijärvi basin was liberated from the ice around 9500 BP (Cal BC 8933–8693; Stuiver & Reimer 1993). According to Seppälä (1980), the retreat of the glacier margin from the Tromso-Lyngen formation to the Kaamasjoki river basin area south west of the Paistunturit area took approximately 500 years. The results of in this investigation indicate that the glacier margin retreated by 60 m per annum at the beginning of the deglaciation and 100 m per annum towards the end. Therefore, the deglaciation of the area took approximately 200–300 years. These results fit well with the estimated timelines presented by Johansson and Seppälä.

Acknowledgements

The author wishes to thank Professor Matti Seppälä for advice, encouragement and constructive suggestions concerning my Master`s thesis, Heikki Hirvas for encouragement, fruithfull discussions and valuable advice concerning the glacial history of Finnish Lapland, Peter Johansson for reviewing and commenting this manuscript, T. Lievonen for assisting in field work in summer 1999 and Tuula and Markku Koskinen for their continuous support.

REFERENCES

Abrahamsson, K.V. 1974. Äkäslompolo-områdets glasialmorfologi och deglaciation. GERUM, Geografiska raportter, Umeå Universitetet, Rapport B: 1, 254 p.

Benn, D.I. & Evans, D.J.A. 1998. Glaciers and Glaciations. Bath: Arnold, 734 p.

Erikstad, L. 1998. Isavsmältningen og israndlinger i Norge. In: Andersen, S. & Pedersen, S.S. (eds.), Israndlinier i Norden. Nordisk Ministerrå, TemaNord 584, Miljø, 97–106.

Eronen, M. & Haila, H. 1990. Highest shoreline and glacial lakes (Map). In: Alalammi, P. (ed.), Atlas of Finland, folio 123–126 Geology. Helsinki: National Board of Survey, Geographical Society of Finland, p 19.

Evans, D.J.A. 2003. Introduction to Glacial Landsystems. In: Evans, D.J.A. (ed.), Glacial Landsystems. London: Arnold, 1–11.

Fitzimons, J.S. 2003. Ice-marginal landsystems: Polar Continental Glacier Margins. In: Evans, D.J.A. (ed.), Glacial Landsystems. London: Arnold, 89–110.

Glasser, N.F., Etienne, J.L., Hambrey, M.J., Davies, J.R., Waters, R.A. & Wilby, P.R. 2004. Glacial meltwater erosion and sedimentation as evidence for multiple glaciations in west Wales. Boreas 33, 224–237.

Heikkinen, O. & Tikkanen, M. 1979. Glacial flutings in northern Finnish Lapland. Fennia 157, 1–12.

Honkala, J. 1982. Deglasiaatiomuodot Peltojokilaaksossa Inarin Lapissa. University of Turku, Department of Geography, Unpublished M.Sc. Thesis, 103 p. [in Finnish]

Hyvärinen, H. & Eronen, M. 1975. Mannerjäätikön sulamisvaiheen aikainen uomasto Koarvikoddsen maastossa Inarissa. Abstract: Glacial drainage channels near Koarrvikodds in Finnish Lapland. Terra 87, 87–94.

Johansson, P. 1988. Deglaciation pattern and ice-dammed lakes along the Saariselkä mountain range in northeastern Finland. Boreas 17, 541–552.

Johansson, P. 1990. Deglasiaatio Saariselän alueella, Itä-Lapissa. University of Turku, Department of Geology, Unpublished Ph.L. Thesis, 92 p. [in Finnish]

Johansson, P. 1994. The subglacially engorged eskers in the Lutto river basin, northeastern Finnish Lapland. In: Warren, W.P. & Croot, D.G. (eds.), Formation and Deformation of Glacial Deposits. Proceedings of the meeting of the Commission on the Formation and Deformation of Glacial Deposits, Dublin, Ireland, May 1991. Rotterdam: A.A. Balkema, 89–94.

Johansson, P. 1995. The deglaciation in the eastern part of the Weichselian ice divide in Finnish Lapland. Geological Survey of Finland, Bulletin 383, 72 p.

Johansson, P. 2005. Moreenipeitteiset jäätikköjokimuodostumat. In: Johansson, P. & Kujansuu, R. (eds.), Pohjois-Suomen maaperä: maaperäkarttojen 1:400 000 selitys (Quaternary deposits of Northern Finland – Explanation to the maps of Quaternary deposits 1:400 000). Espoo: Geologian tutkimuskeskus, 99–105. [in Finnish]

Johansson, P. & Kujansuu, R. 1995. Observations on three subglacial drainage systems (eskers) of different ages in Savukoski, eastern Finnish Lapland. In: Autio, S. (ed.), Geological survey of Finland, Current Research 1993–1994. Geological Survey of Finland, Special Paper 20, 83–93.

Johansson, P., Kujansuu, R. & Mäkinen, K. 1997. Glacial and late-glacial development in NE Kittilä Northern Finland. Geological survey of Finland, Special Paper 23, 59–66.

Kaitanen, V. 1969. A geographical study of the morphogenesis of northern Lapland. Fennia 99, 85 p.

Kaitanen, V. 1989. Relationships between ice-sheet dynamics and bedrock relief on dissected plateau areas in Finnish Lapland north of 69° latitude. Geografiska Annaler 71 A, 1–15.

Karczewski, A. 1975. Morphology and textural-structural features of ground and hummocky moraine in the Paistunturit area, Finnish Lapland. Reports from the Kevo Subarctic Research Station 12, 34–44.

Koskinen, R. 2000. Glasiaalihydrologiset indikaattorit ja mannerjäätikön sulaminen Utsjoen Paistuntureilla. University of Helsinki, Department of Geography, Unpublished M.Sc. Thesis, 106 p.

Kujansuu, R. 1967. On the deglaciation of Western Finnish Lapland. Bulletin de la Commission Géologigue de Finlande 232, 98 p.

Kujansuu, R. 1975. Marrasjärven interstadiaalinen harju Keski-Lapissa. Geologi 27, 45–50. [in Finnish]

Kujansuu, R. 1992. The deglaciation of Finnish Lapland. Geological Survey of Finland, Special Paper 15, 21–31.

Kujansuu, R. & Hyyppä, J. 1995. Vuotson kartta-alueen maaperä. Summary: Quaternary deposits in the Vuotso map-sheet area. Suomen geologinen kartta 1:100 000 lehti 3742, maaperäkarttojen selitykset. Espoo: Geologian tutkimuskeskus, 107 p.

Kujansuu, R., Eriksson, B. & Grönlund, T. 1998. Lake Inarijärvi, northern Finland: sedimentation and late Quaternary evolution. Geologian tutkimuskeskus, Tutkimusraportti 143, 25 p. [in Finnish]

Lundqist, J. 1969. Beskrivning till jordartskarta över jämtlands län. sveriges geologiska undersökning Ca 45, 418 p.

Lundqist, J. 1972. Ice-lake types and deglaciation pattern along the Scandinavian mountain range. Boreas 1, 25–53.

Mannerfelt, C.M. 1945. Några glasialmorfologiska formelement och deras vittnesbörd om inlandisens avsmältningsmekanik i Svensk och Norsk fjällterräng. Geografiska Annaler 27, 396 p.

Mannerfelt, C.M. 1949. Marginal drainage channels as indicators of the gradients of quaternary ice caps. Geografiska Annaler 31, 194–199.

Mansikkaniemi, H. 1967. Geomorphological analysis of Pulmanki-Tana valley in Lapland. Turun yliopiston maantieteen laitoksen julkaisuja 43, 27 p.

Mansikkaniemi, H. 1970. Deposits of sorted material in the Inarijoki-Tana river valley in Lapland. Turun yliopiston maantieteen laitoksen julkaisuja 49, 63 p.

Mansikkaniemi, H. 1973. Morphology of the Utsjoki river valley and the uppermost shoreline on northernmost Finland. Fennia 123, 29 p.

Meriläinen, K. 1976. The granulite complex and adjacent rocks in Lapland, northern Finland. Geological Survey of Finland, Bulletin 281, 129 p.

Mikkola, E. 1932. On the physiography and late-glacial deposits in Northern Lapland. Bulletin de la Commission Géologique de Finlande 96, 88 p.

Nye, J.F. 1973. Water at the bed of glaciers. Symposium on the Hydrology of Glaciers, Cambridge 9–3 sept. 1969. Int. Ass. Scientific Hydrological Publications 95, 189–194.

Penttilä, S. 1963. The deglaciation of the Laanila area, Finnish Lapland. Bulletin de la Commission Géologique de Finlande 203, 71 p.

Piirola, J. 1967. Die glazialen Oberflächenformen und die Entwicklung der Täler auf den Fjelden Marastotunturit and Viipustunturit in Finnisch-Lappland. Annales Academicae Scientiarum Fennicae A III 92, 115 p.

Porola, P. 1981. Deglasiaatio Koarvikoddsin alueella Inarin länsiosassa. University of Turku, Department of Geology, Unpublished M.Sc. Thesis, 102 p. [in Finnish]

Riipinen, P. 1994. Kalliotopografian vaikutus glasiaalihydrologiaan Muotkatuntureiden alueella. University of Helsinki, Department of Geography, Unpublished M.Sc. Thesis, 93 p. [in Finnish]

Price, R.J. 1973. Glacial and fluvioglacial landforms. In: Clay-

ton, K.M. (ed.), Geomorphology Text 5. London: Longman, 95–130.

Seppälä, M. 1971. Evolution of the eolian relief on the Kaamajoki-Kiellajoki river basin in Finnish Lapland. Fennia 104, 88 p.

Seppälä, M. 1980. Deglaciation and glacial lake development in the Kaamasjoki river basin, Finnish Lapland. Boreas 9, 312–319.

Sugden, D.E. & John, B.S. 1976. Glaciers and Landscape. London: Edward Arnold, 361 p.

Shreve, R.L. 1985. Esker characteristics in terms of glacier physics, Katahdin esker system, Maine. Geological Society of America Bulletin 96, 639–646.

Sissons, J.B. 1958. Supposed ice-dammed lakes in Britain, with particular reference to the Eddleston valley, southern Scotland. Geografiska Annaler 40, 159–187.

Stuiver, M. & Reimer, P.J. 1993. Extended 14C data base and revised CALIB 3.0 14C age calibration program, Radiocarbon 35, 215–230.

Sugden, D.E., Denton, G.H. & Marchant, D.R. 1991. Subglacial meltwater channel system and ice sheet overriding, Asgard range, Antarctica. Geografiska Annaler 73 A, 109–121.

Syrilä, S. 1964. Retreat of continental ice and fluvioglacial erosion features on the Tuolbanjaugoaivi Fjeld in northernmost Finland. Ann. Univ. Turku A: 32, 338–345.

Syrilä, S. 1970. Mannerjään sulaminen Akujoen-Fiellujoen laaksossa Utsjoella. University of Turku, Department of Geography, Unpublished M.Sc. Thesis, 59 p. [in Finnish]

Tanner, V. 1915. Studier öfver kvartärsystemet i Fennoskandias nordliga delar III. Fennia 36, 815 p.

Tanner, V. 1938. Die Oberflächengestaltung Finnlands. Bidrag till kännedom av Finlands natur och folk 86, 762 p.

Quaternary studies in the northern and Arctic regions of Finland
Edited by Antti E.K. Ojala
Geological Survey of Finland, Special Paper 40, 53– 58, 2005.

TIMING OF EARLY HOLOCENE LANDSLIDES IN KITTILÄ, FINNISH LAPLAND

by
Raimo Sutinen

Geological Survey of Finland, P. O. Box 77, FIN-96101, Rovaniemi, FINLAND
E-mail: raimo.sutinen@gtk.fi

Sutinen, Raimo, 2005. Timing of early Holocene landslides in Kittilä, Finnish Lapland. *Geological Survey of Finland, Special Paper 40*, 53–58, 6 figures.

Northern Fennoscandia has experienced high-magnitude (>Mag 7) neotectonic fault deformations that are attributable to intra-plate glacio-isostatic rebound. According to this conceptual model, landslides have been linked to these postglacial seismic events. The age of the landslide features, however, has remained obscure. Two landslide sites (Hanhilaki and Pyhävaara) in the vicinity of the Suasselkä (-Homevaara) fault line in Kittilä, western Finnish Lapland were investigated for texture, fabric and age. The sites differed with respect to clay fraction content, but no evidence was found to indicate that clayey Tertiary weathering products to have contributed to sliding mechanisms. Fabric was either non-oriented (Pyhävaara) or preferred to slide direction (Hanhilaki), but was inconsistent with known ice-flow phases. Buried fragments of subfossil downy birch (*Betula pubescens*) trunks were found beneath the landslide materials at the Hanhilaki site, and the sampled wood yielded ^{14}C-age of 8,720±170 yrs B.P. (9,730 cal B.P.) The age of Pyhävaara slide still remained unclear. The present study demonstrated that only a part of landslides in Lapland are formed in the proximity of retreating continental ice, but a substantial number of slides occurred during the next 2,000 years after the ice disappearance. Evidently the early Holocene landslides, triggered by seismic tremors, were associated with inter-seasonal soil saturation, but with snowmelt as a major contributor to sediment instability.

Key words (GeoRef Thesaurus, AGI): landslides, neotectonics, fabric, textures, plant macrofossils, Betula pubescens, C-14, Holocene, Kittilä, Finland

Introduction

Landslides are common features in northern Fennoscandia, and those have been linked to seismic events in association with postglacial rebound (Kujansuu 1972, Lagerbäck 1990, Dehls et al. 2000) (Figs. 2–3). Even though lithostratigraphic evidence suggests neotectonic faulting to be generated during late-glacial or post-glacial time period (Lagerbäck 1979), the age of the landslides has remained obscure.

A number of SW-NE trending neotectonic fault lines have been recognized in northern Fennoscandia, among others Suasselkä and Pasmajärvi faults in Finland, Pärvie fault in Sweden as well as Stuorragurra fault in Norway (Kujansuu 1964, Lundqvist & Lagerbäck 1976, Lagerbäck 1979, Dehls et al. 2000) (Fig. 1). Neotectonic deformations have been associated with glacio-isostatic rebound (Dehls et al. 2000, Olesen et al. 2004, Ojala et al. 2004), but even though the high-magnitude (>Mag 7) seismic tremors evidently

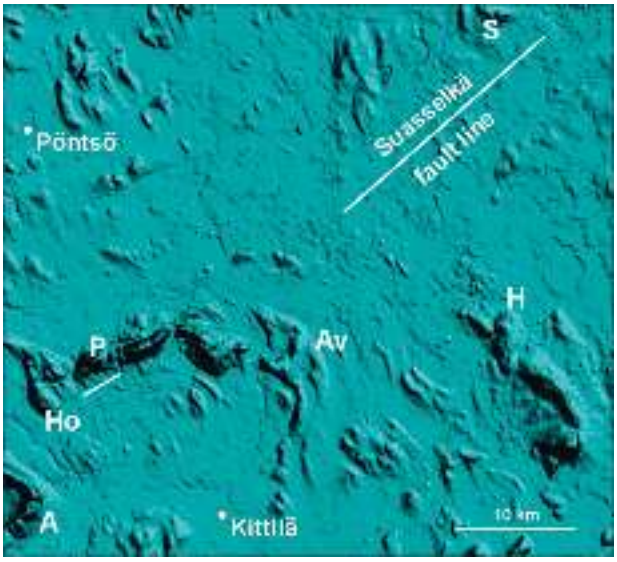


Fig. 1. Study sites in Kittilä, Finnish Lapland. H=Hanhilaki, P=Pyhävaara, S=Suasselkä, Ho=Homevaara, A=Aakenustunturi, Av=Akanvalkko, Thick line shows Suasselkä-Homevaara neotectonic fault line. (Details for A and Av, see Kujansuu 1972)

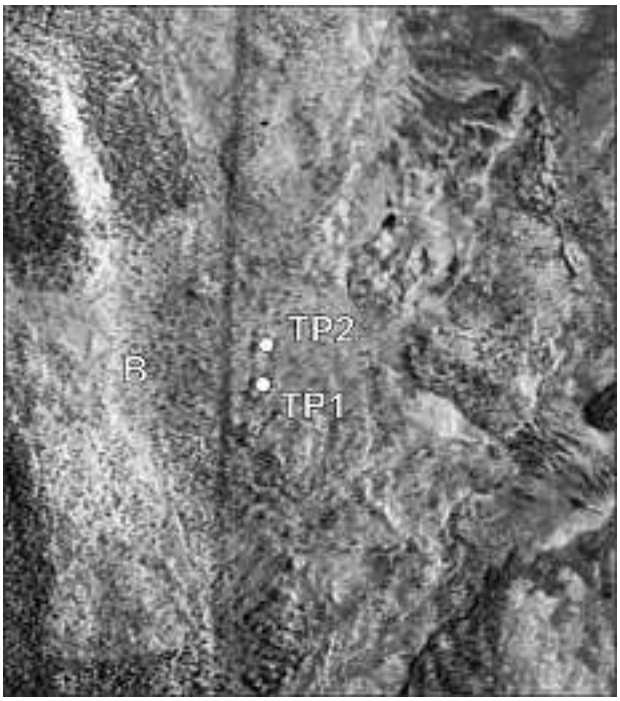


Fig. 2. Airphoto showing Hanhilaki landslide features in Kittilä, Finnish Lapland. B=Backwall, TP1=Test Pit 1, TP2=Test Pit 2. Size of the slide scar 500 by 1200 m.

occurred soon after deglaciation 9,000–10,000 years ago (Lagerbäck 1979), the age, however, has been inconclusive in the intra-plate Fennoscandia. Offshore sediments in the Norwegian fjords as well as lake sediments, however, provide evidence about the seismic events, dated with the aid of radiocarbon ages of the marine fauna and gyttja sediments. Until now no dating has been presented for any of the landslides in Finnish Lapland (cf. Kujansuu 1972, Lagerbäck 1990). The objective of this study was to find out i) if the slides were attributed to texture of slide materials, e.g. clayey Tertiary weathering products initially susceptible for sliding and ii) to discuss the age and timing of the slides in western Finnish Lapland.

Materials and methods

Study sites

Stereoscopic airphoto interpretation and digital elevation model (DEM by the National Land Survey) were applied to recognize landslides. For field surveys the sites selected on the basis of the morphological features (slide scars and accumulation forms) and location lie within the 500-m forest logging roads. Easy access allowed tractor trenching/ test pits through the slide materials, originally glacial tills of undefined stratigraphic position. Two sites targeted for test pits and trenching: Hanhilaki (3435136, 7525372, 233 m a.s.l., Fig. 2) and Pyhävaara (3400684; 7521243; 240 m a.s.l.) (Fig. 3). Both sites are located in the vicinity of the Suasselkä fault, previously reported by Kujansuu (1972), such that Hanhilaki is located 17 km SE from the Suaselkä fault and Pyhävaara site is in the immediate proximity of the Homevaara (associate with Suasselkä) fault in Kittilä, northern Finland (Fig. 1). The slopes were as follows, ca. 15% at Pyhävaara site and 10–12% at Hanhilaki site, respectively. The size of the slide scars varied from 300 by 500 m at Pyhävaara (Fig. 3) to 500 by 1200 m at Hanhilaki (Fig. 2). The back-wall of the scar at Pyhävaara site is approximately 10 m demonstrating a rotational NW-NE-oriented slide. At Hanhilaki it appears that much of the slope of the hill had been slided roughly to the west (Fig 2). The slided material was originally till in both cases. These seismically induced slide materials, ‘seismites’, were piled up as mounds and ridges morphologically similar to those of hummocky moraines (Figs. 2–4). The landslide hummocks varied from 4 to 6 m in height. Test pits and trenches were excavated for fabric analyses (Figs. 5–6), for sampling and particle-size analyses with sieving and conventional sedimentation (aerometer), and for sampling of buried organic material for radiocarbon dating in the Radiocarbon Laboratory at the Geological Survey of Finland.

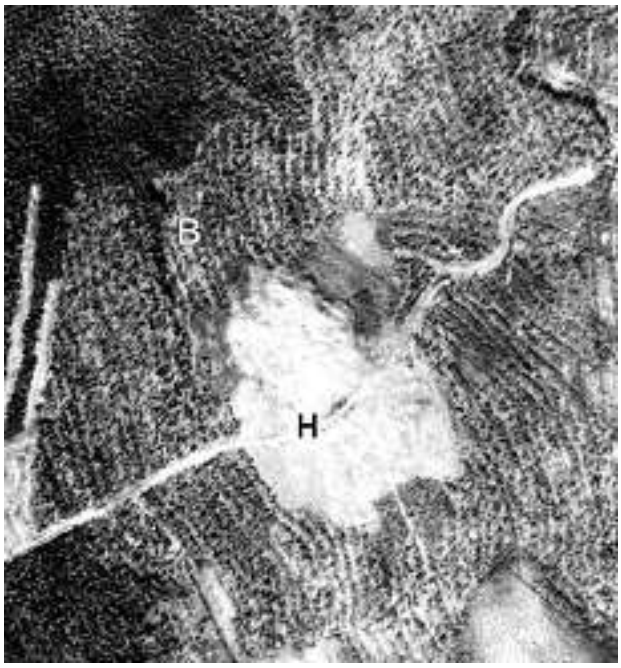


Fig. 3. Airphoto showing Pyhävaara landslide features in Kittilä, Finnish Lapland. B=Backwall, H=Landslide Hummocks. Size of the slide scar 300 by 500 m.

Fig. 4. Outcrop of the Pyhävaara 'seismite' hummock in Kittilä, Finnish Lapland. (Photo: R. Sutinen, 1985)

Results and discussion

Fabric of landslide 'seismites'

Fabric analyses made from the walls of the test pits revealed that landslide materials, 'seismites', could not to be connected to the Weichselian ice-flow phases in the Kittilä area. On the basis of glacial dispersal patterns of magnetite-bearing stones at Suasselkä (location in Fig. 1) three ice-flow phases, NW-SE (Early Weichselian), N-S (Mid-Weichselian) and SW-NE (Late Weichselian) trending phases have been identified (Sutinen 1992). Particularly the Pyhävaara slide (Fig. 4) exhibited non-oriented pattern of fabric in the hummocks (Fig. 5). Hanhilehto 'seismites' had preferred orientation (Fig. 6) that appeared to be sediment flow indicator rather than attributed to the ice-flow patterns. This is coincidental with the observations by Kujansuu (1972) indicating the orientation of the slided materials to deviate from that of adjacent 'non-slided' tills. However, it is not known if large-magnitude (>Mag 7) earthquakes (Lagerbäck 1990, Dehls et al. 2000) are able to impact on the primary glacigenic fabric of tills. Lagerbäck (1990) for example presented vertical grading of till due to seismicity in northern Sweden. However, low magnitude (<Mag 3.8) seismic event had not impacted on till fabric beyond the landslide scar as observed by Sutinen (1992) at the margin of Burroughs Glacier in Alaska. Except tills, primary structures of waterlain sediments are known to dramatically deform during the seismic events (Lagerbäck 1990, Menzies & Taylor 2003, Bøe et al. 2004).

Texture of landslide 'seismites'

Particle-size analyses revealed that fine fraction contents (<60 μm) were rather similar at both sites, 32–46% for six samples from Hanhlaki and 34–49% for two samples from Pyhävaara, respectively. Clay fraction content (<2 μm), however, was significantly higher (6–11%) for Hanhilaki samples as compared to those for Pyhävaara (3–4%), respectively. The matrix of Pyhävaara 'seismite' was similar to those in tills from the landslide scars elsewhere in northern Finland and Sweden as presented previously by Kujansuu (1972) and Lagerbäck (1990). No evidence was found to indicate soil sequences with higher clay content to have acted as 'gliding base' at the Pyhävaara site. In the case of Hanhilaki site texture was not consistently higher in deeper sequences, either. Clay fraction content in Pit 1 was as follows: 8.9% at 4.5-m-depth, 6.3% at 3-m-depth, and 6.7% at 1.2-m-depth, but in Pit 2 as follows: 11.4% at 3-m-depth and 10.8% at 1.5-m-depth, respectively.

Buried subfossils

Buried organic material was found as a dark grey intermixture with clayey 'seismite' at Hanhilaki site, not at Pyhävaara. No signs of *in situ* soil profile (Spodosol) were discernible in the excavated trenches and the observed plant remnants were incorporated into 'seismite' to 3–4.5-m-depth. Evidently the slide event had deformed the original stratigraphy of the

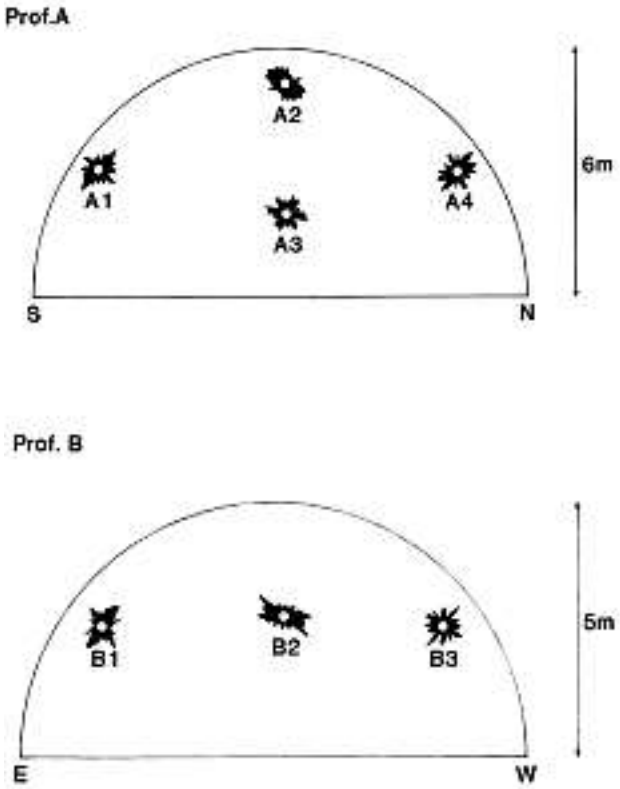


Fig. 5. Fabric of two Pyhävaara 'seismite' hummocks in Kittilä, Finnish Lapland.

till sequences. Among well-preserved leaves of trees and shrubs, moss fragments as well as twigs of *Betula* ssp., also fragments of subfossil tree trunks were found. The sampled wood, even though decomposed, was recognized as *Betula pubescens* ssp. *pubescens* or *tortuosa* and yielded ^{14}C-age of 8,720±170 yrs B.P. (9,730 cal B.P.) (Su-1517).

The ^{14}C-age (8,720±170 yrs B.P.) of the buried birch at Hanhilaki provides firm macrofossil evidence indicating that birch migrated Kittilä region as early as during the Preboreal period. The age of subfossil downy birch is also synchronistic with the age range from 8,140±180 yrs B.P. to 8,280±130 yrs B.P. given for peat deposits overlying the landslide scars elsewhere in northern Finland and Sweden (Kujansuu 1972, Lagerbäck 1990). However, one of the datings of the overlying peat in Kujansuu (1972) yielded 6,610±175 yrs B.P. This site, Aakenustunturi (shown in Fig. 1), is located only 45 km WWS from Hanhilaki, hence it is not likely that these two landslides were attributed to retreating margin of continental ice. Rather the age-gap of >2,000 years is the minimum timeframe for the seismic activity that had contributed to landslides in Kittilä.

The pre-landslide presence of birch forest at Hanhilaki signifies that the slide event was not ice-marginal feature. However, Akanvalkko site (location shown in Fig. 1), only 18 km west of Hanhilaki, has been indicated as occurred at the margin of the retreating ice (Kujansuu 1972). Therefore, the postglacial faults may have been active enough to trigger landslides for more than 2,000 years since large-magnitude event (earthquake >Mag 7) soon after deglaciation 9,000–10,000 years ago (Lagerbäck 1979, 1990). The reactivated fault zones may have faced several seismic deformations to be able to trigger landslides within a time span of couple of years to several hundreds

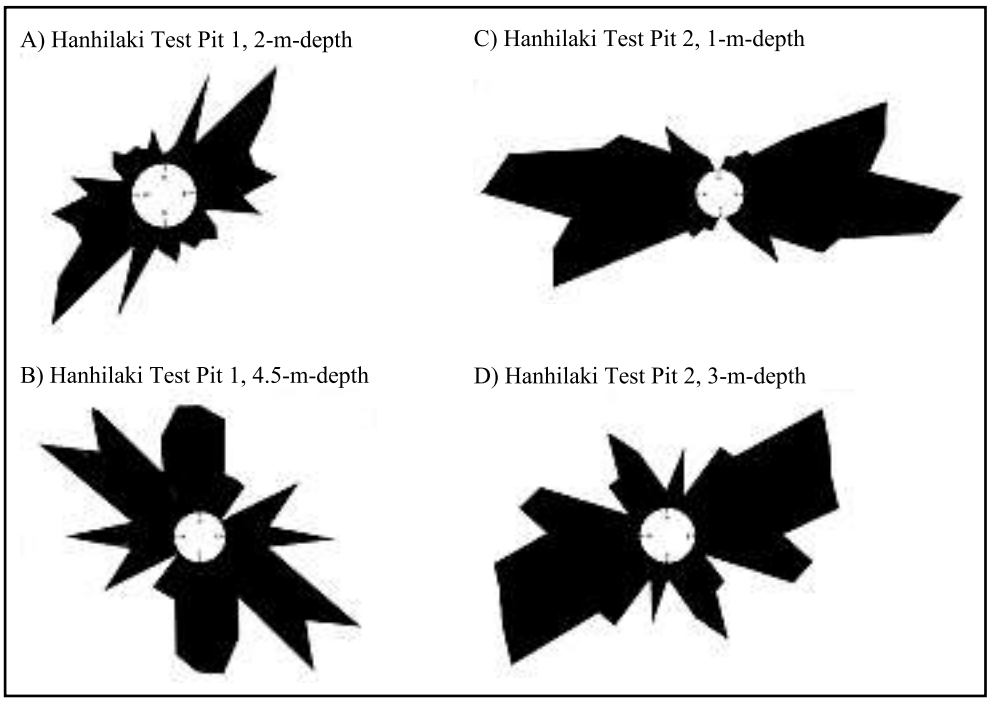


Fig. 6. Fabric of two Test Pits excavated into the Hanhilaki 'seismite' in Kittilä, Finnish Lapland.

of years or even thousands of years (Sutinen 1990, 1992, Dehls et al. 2000, Ojala et al. 2004, Olesen et al. 2004). It has been indicated that postglacial rebound of app. 4 mm/yr associated with earthquakes (<Mag 4.4) in Glacier Bay area of SE Alaska had contributed to several ice-marginal landslides within few years in the 1980's (Sutinen 1992). In addition, sediment mass flows triggered by large-magnitude earthquakes in a number of fjords in western and northern Norway date back to 11,700–9,000 yrs B.P. (Bøe et al. 2004, Olesen et al. 2004). Hence the landslide activity had time-transgressively attributed to glacio-isostatic rebound following retreating pattern of the continental ice in Fennoscandia. It is likely that landslides in Kittilä were triggered by seismic events of the neotectonic Suasselkä (-Homevaara) fault. The number of seismic events, however, has yet to be studied e.g. with varve-chronology of the lake sediments (Bøe et al. 2004).

Slide susceptibility

Studies on landslide susceptibility at the margin of Matanuska Glacier in Alaska have demonstrated that excess soil water content (θ_v) is a prerequisite factor for the sediment instability (Hänninen & Sutinen 1994). On the basis of four-step linear dielectric (ε) mixing model, geotechnical resistance will significantly decrease at full saturation of sediments, i.e. ε=30 ($\theta_v \approx 40\%$), but beyond this critical point sediments are highly susceptible for sliding (Hänninen & Sutinen 1994). In the case of Hanhilaki, covered by forest dominated by downy birch, tills had not influenced by melting ice, but the θ_v of tills followed intra- and inter-seasonal climatic events prevailing during the early Holocene. Since the annual precipitation was not lower at the time of landslides as compared to present climatic conditions (Bjune et al. 2004), one would assume that most obvious timing for the slope instability was late spring-early summer. Inter-seasonal dielectric monitoring of the soil θ_v has indicated that the variation magnitudes of the soil θ_v are site and texture specific, but with snowmelt as a major contributor to sediment instability (Hänninen 1997, Sutinen et al. 1997). Soil saturation ε>30 (θ_v>40%) of silty tills may last 2–8 weeks after disappearance of snow cover in Lapland, but that of sandy tills is either totally lacking or is a matter of only few days (Sutinen et al. 1997). The slope (≈15%) at Pyhävaara site was not much steeper as compared to that (10–12%) at Hanhilaki site, hence the timing of the seismic event was of a great importance with respect to landslides, particularly in the case of sandy tills (see Kujansuu 1972, Lagerbäck 1990) (Fig. 3). The findings at Hanhilaki and Pyhävaara sites demonstrate that texture was not a primary factor for slide susceptibility, but rather the slope stability was controlled by the seasonal variations in soil water content (Hänninen & Sutinen 1994, Sutinen et al. 1997). It appears likely that part of the landslides were ice-marginal features (Kujansuu 1972, Sutinen 1990, 1992), but a substantial number of slides had occurred during the time-span of more than 2,000 years after the ice disappearance, and had resulted from coincidental snowmelt saturations and triggering seismic tremors in Kittilä, Finnish Lapland.

Acknowledements

Material for the present paper was obtained during mapping of Quaternary deposits by the Geological Survey of Finland. I thank Liisa Ikonen for identifying wood material, Reijo Lampela for scanning the photos and Viena Arvola for editing work.

REFERENCES

Bjune, A.E., Birks, H.J.B. & Seppä, H. 2004. Holocene vegetation and climatic history on a continental-oceanic transect in northern Fennoscandia based on pollen and plant macrofossils. Boreas 33, 211–223.

Bøe, R., Longva, O., Lepland, A., Blikra, L.H., Sønstegaard, E., Haflidason, H., Bryn, P. & Lien, R. 2004. Postglacial mass movements and their causes in flords and lakes in western Norway. Norwegian Journal of Geology 84, 35–55.

Dehls, J.F., Olesen, O., Olsen, L. & Blikra, L.H. 2000. Neotectonic faulting in northern Norway, the Stuoragurra and Nordmannvikdalen faults. Quaternary Science Reviews 19, 1447–1460.

Hänninen, P. 1997. Dielectric coefficient surveying for overburden classification. Geological Survey of Finland, Bulletin 396, 72 p.

Hänninen, P. & Sutinen, R. 1994. Dielectric prediction of landslide susceptibility: A model applied to recent sediment flow deposits. In: Oliveira, R., Rodrigues, L.F., Coelho, A.G. & Punha, A.P. (eds.), Proc. 7th International IAEG Congress. Rotterdam: Balkema, 137–144.

Kujansuu, R. 1964. Nuorista siirroksista Lapissa (Summary: Recent faults in Lapland). Geologi 16, 30–36. [in Finnish]

Kujansuu, R. 1972. On landslides in Finnish Lapland. Geological Survey of Finland, Bulletin 256, 1–22.

Lagerbäck, R. 1979. Neotectonic structures in northern Sweden. Geologiska Föreningens i Stockholm Förhandlingar 100, 263–269.

Lagerbäck, R. 1990. Late Quaternary faulting and paleoseismicity in northern Fennoscandia with particular reference to the Landsjärv area, northern Sweden. Geologiska Föreningens I Stockholm Förhandlingar 112, 333–354.

Lundqvist, J. & Lagerbäck, R. 1976. The Pärve Fault: A late-glacial fault in the Precambrian of Swedish Lapland. Geologiska Föreningens i Stockholm Förhandlingar 98, 45–51.

Menzies, J. & Taylor, J. 2003. Seismically induced soft-sediment microstructures (seismites) from Meikleour, western Stratmore, Scotland. Boreas 32, 314–327.

Ojala, V.J., Kuivamäki, A. & Vuorela, P. 2004. Postglacial deformation of bedrock in Finland. Geological Survey of Finland, Nuclear Waster Disposal Research, Report YST-120, 23 p.

Olesen, O., Blikra, L.H., Braathen, A., Dehls, J.F., Olsen, L., Rise, L., Roberts, D., Riis, F., Faleide, J.I. & Anda, E. 2004. Neotectonic deformation in Norway and its implications: a rewiew. Norwegian Journal of Geology 84, 3–34.

Sutinen, R. 1990. Landslides near Burroughs Glacier, Southeast Alaska. Geonytt 17, 110.

Sutinen, R. 1992. Glacial deposits, their electrical properties and surveying by image interpretation and ground penetrating radar. Geological Survey of Finland, Bulletin 359, 123 p.

Sutinen, R., Hänninen, P., Mäkitalo, K., Penttinen, S. & Sutinen, M.-L. 1997. Snowmelt saturation restricts Scots pine growth on fine-grained tills in Lapland. CRREL Special Report 97–10, 507–512.

Quaternary studies in the northern and Arctic regions of Finland
Edited by Antti E.K. Ojala
Geological Survey of Finland, Special Paper 40, 59– 65, 2005.

LANDFORM DEVELOPMENT DURING THE WEICHSELIAN GLACIATION IN PERÄPOHJOLA, FINLAND

by
Pertti Sarala

Geological Survey of Finland, P.O. Box 77, FIN-96101, Rovaniemi, FINLAND
E-mail: pertti.sarala@gtk.fi

Sarala, Pertti, 2005. Landform development during the Weichselian glaciation in Peräpohjola, Finland. *Geological Survey of Finland, Special Paper 40*, 59–65, 6 figures.

Glacial morphology, stratigraphy, dynamics and glacial history were investigated in the area of Peräpohjola, southern Finnish Lapland. Particularly, the areas of active ice streams and passive interlobes were distinguished with the help of glacial morphological interpretations. The active ice streams were depicted as having a glacial morphology composed of ribbed moraines, drumlins and flutings. In the interlobate area, cover and ground moraines with no indication of glacial flow or morphology of younger glacial phase were common.

As a result, two models of landform and till stratigraphical development are presented. In the first model, two glacial phases occurred, the first one of which was during the Early Weichselian. The second model includes a new interpretation where the first larger glaciation did not exist until the Middle Weichselian and the ice-free interstadial that followed occurred before the Late Weichselian glaciation.

Key words (GeoRef Thesaurus, AGI): glacial geology, glaciation, landform evolution, moraines, till, stratigraphy, Pleistocene, Weichselian, Peräpohjola, Finland

Introduction

The area of Peräpohjola in southern Finnish Lapland (Fig. 1) was frequently situated in the central area of former glaciations. Due to the close vicinity of the glacier's centre, the effect of glacial erosion has varied greatly in places. The glacial drift sheet is mostly thin, especially in the areas of active ice streams, where glacial erosion has been strong. For that reason, older glaciogenic deposits, for example of Saalian age or older, were commonly eroded away. However, under passive, non-eroding ice lobes, older deposits and even morphology could have been preserved (cf. Kleman 1994).

Glacial morphology, dynamics and history were investigated as a joint project with the Geological Survey of Finland (GTK) and the University of Oulu from 1992–1996. Regional geomorphology was studied using aerial photo interpretation and elevation models. In case studies, test pit excavations were used to clarify composition, structure and stratigraphy of moraine formations. The analysis of data continued during the present decade at the GTK. Particular interest was paid to the application of knowledge of

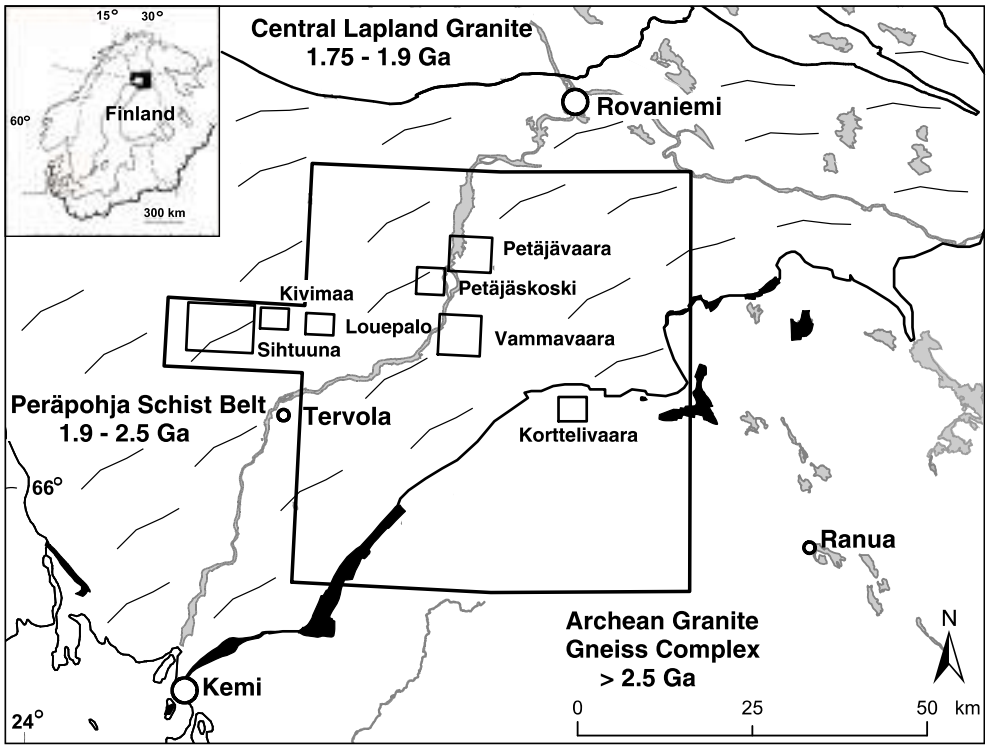


Fig. 1. The location of the study area and test sites together with a generalized bedrock map of the Peräpohja Schist Belt in Peräpohjola, Finland. The area is situated in the centre of Late Weichselian glaciation. (Topographic features © National Land Survey of Finland, permission number 63/LA/05)

Quaternary deposits and processes in ore prospecting, because many Au-Cu rich boulders and metal anomalies were observed in till. A detailed investigation was done in seven test sites in the area between Rovaniemi and Tervola.

Glacial morphology and till stratigraphy

The glacial morphology of the area (Fig. 2) was formed during two glacial phases. An older glacial advance was depicted as a drumlin field deposited during the glacial flow from the north-northwest to the south-southeast (cf. Sutinen 1992). The following melting phase is also seen as a series of marginal formations from the southern side of Pudasjärvi to the northwest (Mäkinen 1985). These formations are mainly till-covered as a result of a later glacial advance which flow direction has generally been towards the east. This glaciation was represented as an association of active-ice landforms from hummocky moraines and ribbed moraines to drumlins and flutings (Aario 1977, Sarala et al. 1998). In the east, in the areas of Kuusamo and Posio, streamlined moraines form the large Kuusamo drumlin field. When moving to the west, ribbed moraines transversal to the latest glacial flow are dominant in the area of Peräpohjola. Three large and quite uniform ribbed moraine fields exist in the areas of Tervola and Ranua and also on the western side of Kemijärvi. All of those landform types are indicative of active ice streams but between the ice streams is the area of the passive Ranua lobe, where cover and ground moraines are common without special morphology or clear glacial flow indication.

Ribbed moraines consist of three subtypes in the area; hummocky ribbed moraines, Rogen moraines and minor ribbed moraines (Fig. 3) (cf. Hättestrand 1997, Sarala 2003). The Rogen moraine is the only type that clearly has streamlined features, for example drumlinized or fluted features superimposing the ridges (cf. Lundqvist 1969). A convergent feature of all ribbed moraines is their formation under the fast flowing glacier and in the boundary of the cold- and warm-based ice sheet during the early stage of deglaciation (Hättestrand 1997).

The composition, structures and till stratigraphy of ribbed moraines in Peräpohjola proved that the ridges were formed in a two-step process (Sarala 2005). During the first stage, a brittle mass of pre-existing sediments and the bottommost part of the ice sheet were fragmented due to tension seen as featherlike cracks under compressive ice flow. After the transition of fragmented blocks, a rib-like morphology formed. During the second phase, after the tension decreased, moving ice transported and re-deposited pre-existing

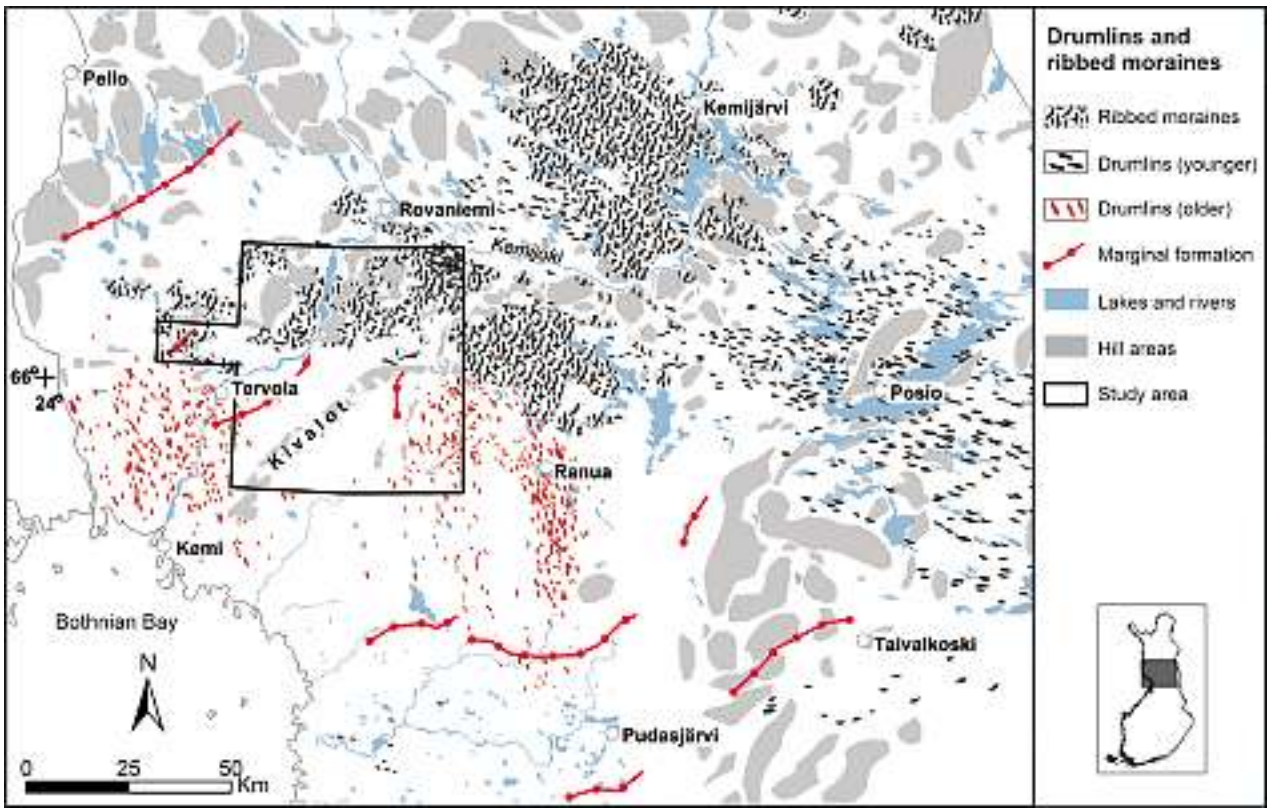


Fig. 2. The occurrence of drumlins, ribbed moraines and older marginal formations in southern Finnish Lapland. The older drumlin field was preserved under the cold-based, passive Ranua interlobate area. (Topographic features © National Land Survey of Finland, permission number 63/LA/05)

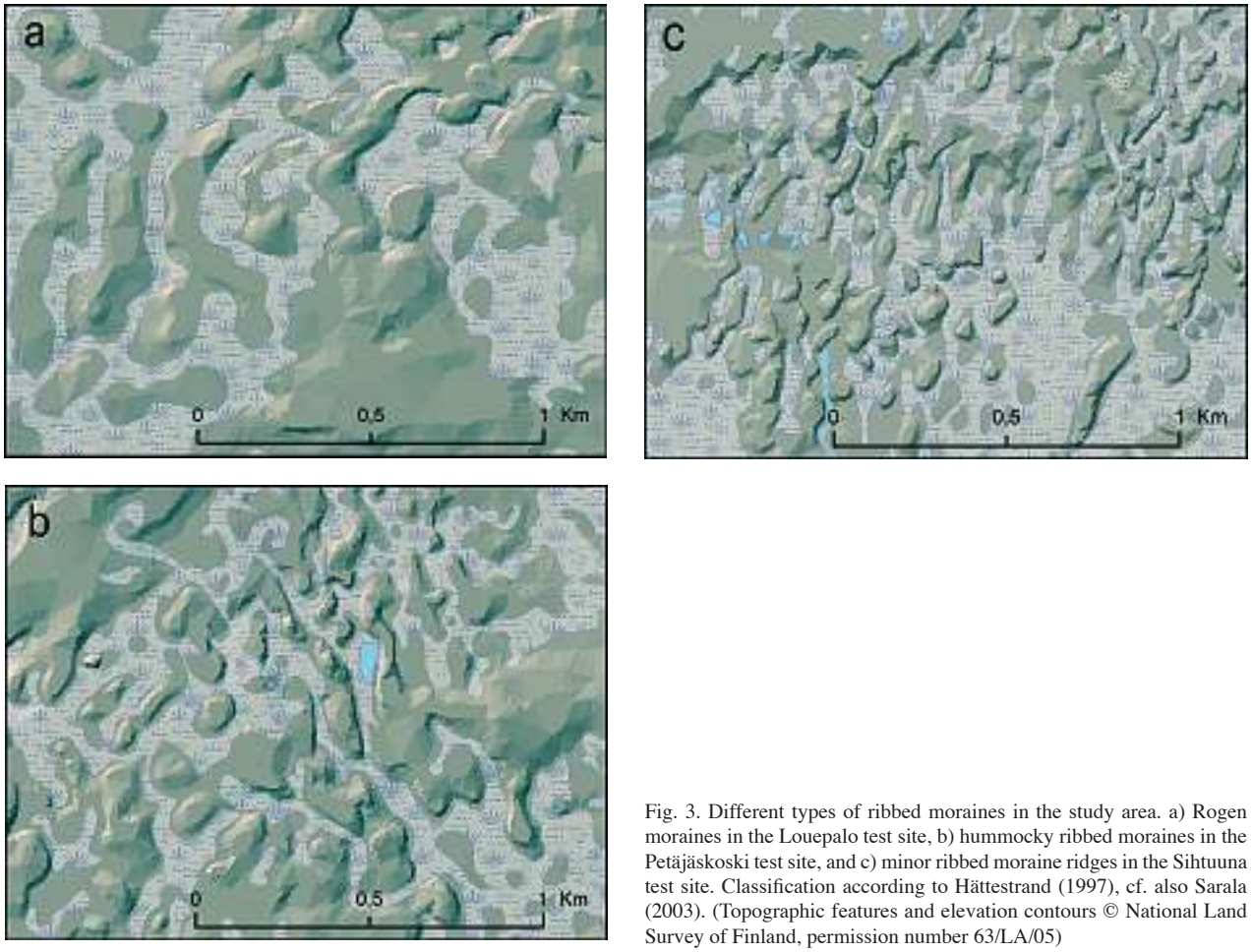


Fig. 3. Different types of ribbed moraines in the study area. a) Rogen moraines in the Louepalo test site, b) hummocky ribbed moraines in the Petäjäskoski test site, and c) minor ribbed moraine ridges in the Sihtuuna test site. Classification according to Hättestrand (1997), cf. also Sarala (2003). (Topographic features and elevation contours © National Land Survey of Finland, permission number 63/LA/05)

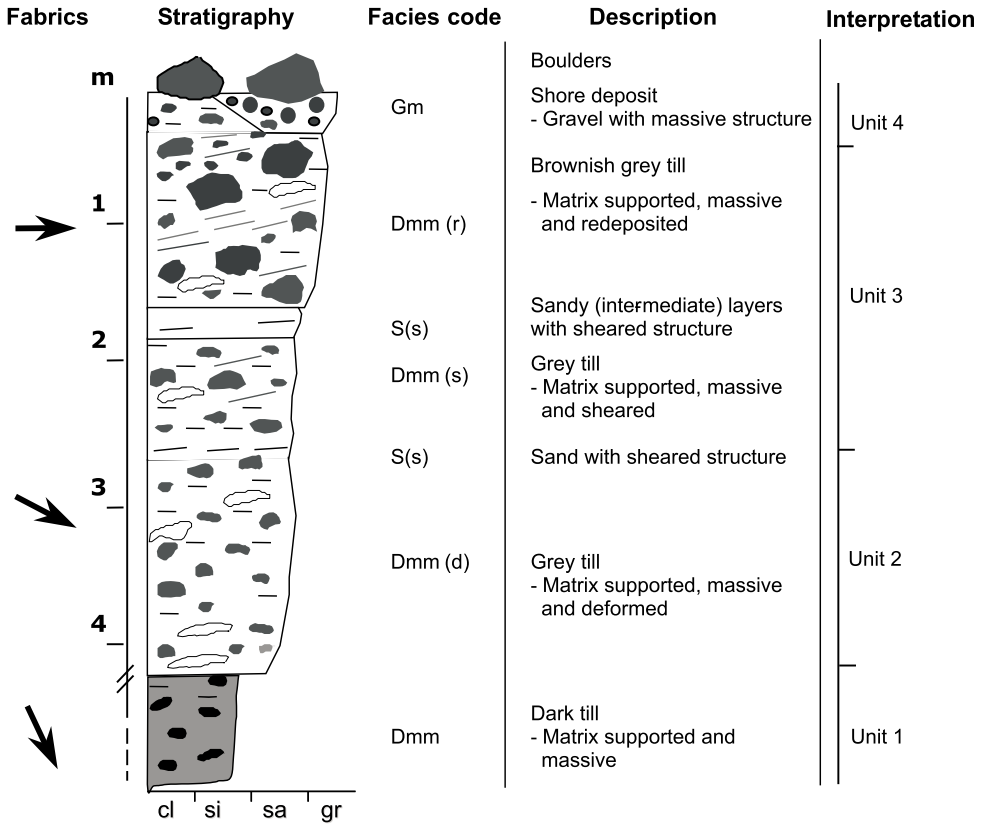


Fig. 4. Composition stratigraphy of ribbed moraines in the study area showing various lithofacies and their interpretation. Arrows depict mean glacial flow direction in different till units.

tills and sediments from the intermediate areas to the surface of ridges as a result of the freeze-thaw process. As glacial movement continued, the bedrock surface was also quarried and boulders, which provided a good indication of local bedrock, were deposited on the surface of ribbed moraine ridges. Thus, the bouldery surface is a characteristic feature for all ribbed moraines in the Peräpohjola area. Boulder fans are short, only some tens or hundreds of metres long, and they were proven to be a useful feature, for example, in ore prospecting or bedrock mapping in the area (cf. Sarala et al. 1998, Sarala & Rossi 2000).

During test pit excavations, ribbed moraine ridges were found to be composed of four till units (1–4, Fig. 4). The lowest till unit (1) was composed of compact, dark grey till, deposited under glacial flow towards the south-southeast. The grey, massive till unit (2) above that represented the advancing glacial phase with a flow direction from the west-northwest. The third, brownish grey, sandy or gravelly till unit was deposited during the early stage of deglaciation, that is during the second phase of ribbed moraine formation presented earlier. The third till unit is composed of redeposited material and is a mixture of pre-existing tills or sediments, englacial material and fresh rock fragments quarried from the surface of local bedrock. The glacial flow direction was from the west. During the final melting phase at the end of deglaciation, the uppermost, brownish grey and heterogeneous till unit was deposited (4). It can be distinguished only in certain places, because it has been reworked during the post-glacial Ancylus Lake and is usually difficult to distinguish from the shore deposits.

The tills presented above correlate with the Weichselian glaciation (e.g. Kujansuu 1967, Korpela 1969, Hirvas 1991, Sutinen 1992). In other studies published by Mäkinen (1979) and Sutinen (1992), the fifth till unit, interpreted to be of Saalian glaciation, was observed under the Weichselian till units.

Landform development

The present study confirmed the opinion that the Peräpohjola area in southern Finnish Lapland was glaciated only twice during the Weichselian glaciation. During the first glacial phase, dark grey till, older drumlin field and marginal deposits were deposited (cf. Mäkinen 1985, Sutinen 1992). That phase was followed by the warmer, ice-free phase, when *Betula* dominant tundra vegetation spread to northern Finland.

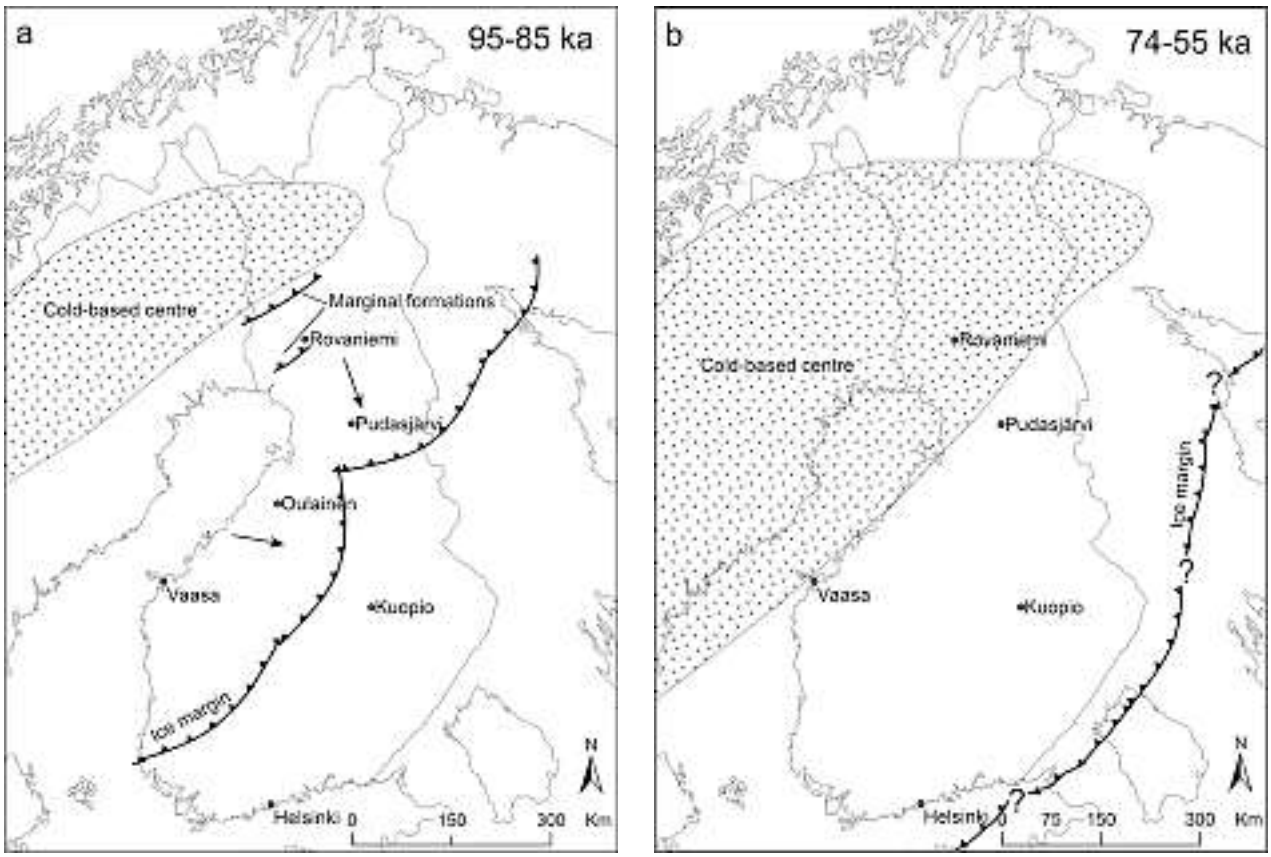


Fig. 5. The first model of landform development during the Weichselian in southern Finnish Lapland. a) Phase I - the Early Weichselian glaciation advanced to the level of Oulainen and Pudasjärvi. Glacial flow directions of ice streams are depicted as arrows. b) Phase II - the Middle Weichselian glaciation advanced at least to the level of southern Finland (cf. Svendsen et al. 2004) and may have continued to the end of Late Weichselian without a break.

This phase is seen as inter-till organic deposits between the till units 1 and 2 that Korpela (1969) described as the Peräpohjola Interstadial deposit. The warmer interstadial terminated at the second Weichselian glacial phase. The glacier advanced over wide areas, maybe as far as the Weichselian maximum during that phase. At its largest stage, core areas were largely cold-based and an ice divide existed above the present Bothnian Bay, where it migrated to the northwest as deglaciation progressed. During the melting phase, the outer zone of the glacier was divided into lobes of active ice streams and passive areas. Active-ice moraine formations, such as ribbed moraines, drumlins and flutings, are common in the areas of ice streams, while cover and ground moraines or older morphology are distinguished in the passive interlobate area.

The timing of Weichselian glacial phases is difficult to estimate because of the small amount of determinations and the lack of reliable dates in southern Finnish Lapland. However, the comparison of stratigraphical and morphological development in the area with the present knowledge of timing presented in literature (Korpela 1969, Kujansuu 1975, Siivonen 1975, Mäkinen 1979, Punning & Raukas 1983, Hütt et al. 1984, Hirvas 1991, Mäkinen 1999, Helmens et al. 2000, Mäkinen in press) can be summarized as two models:

Model 1

According this model, the first glacial phase existed in the middle of the Early Weichselian (95–85 ka ago) (Fig. 5), when the glacier advanced as separate ice streams to the Ostrobothnian and Kainuu areas (cf. Lundqvist 1992). The Peräpohjola Interstadial followed that phase during the Odderade (85–74 ka ago) (cf. Nenonen 1995, Lunkka et al. 2004). At the beginning of Middle Weichselian, the climate became cold once again and the second glacial phase began. Since the organic intermediate layers were observed only in one interstadial phase in Lapland, the second glacial phase must have lasted until the end of the Late Weichselian without oscillation of the glacier's margin to the southern Finnish Lapland level. However, the stratigraphical and dating evidence does not agree with this interpretation.

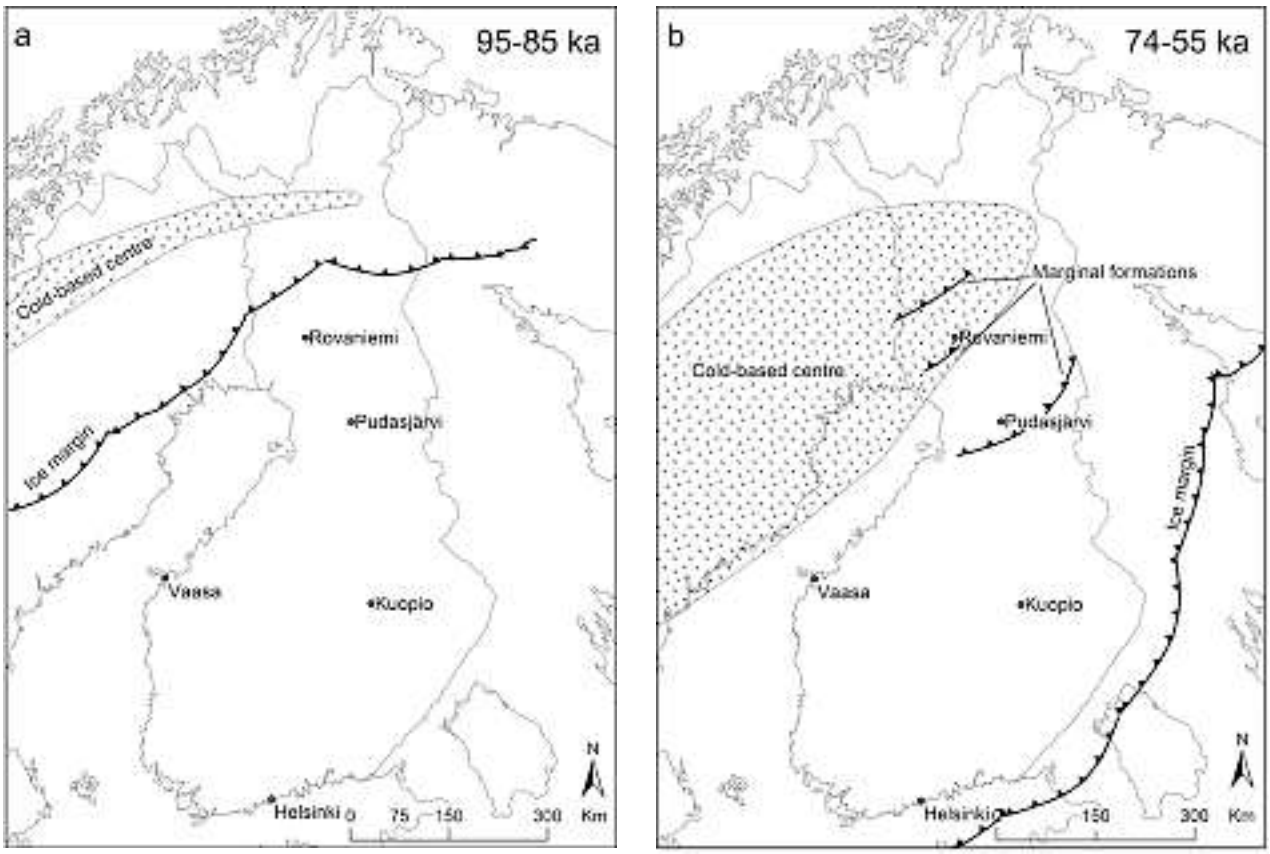


Fig. 6. The second model of landform development during the Weichselian in southern Finnish Lapland. a) Phase I - the Early Weichselian glaciation covered only the areas of central Lapland while southern Finnish Lapland stayed ice-free. b) Phase II - the Middle Weichselian glaciation covered the area of Peräpohjola, advancing at least to the level of southern Finland (cf. Svendsen et al. 2004). Marginal formations in northern Finland were formed during the first deglaciation. The interstadial followed it in the oxygen isotope stage 3 before the Late Weichselian glaciation.

Model 2

In the second model, almost all of Finland was ice-free except central and northern Lapland during the Early Weichselian (Fig. 6). The climate was quite cool and due to glacial isostacy, the ground was about 90–100 m higher than present at the end of the Early Weichselian (cf. Nenonen 1995).

The first larger glacial phase did not existed until the beginning of the Middle Weichselian (74–55 ka ago). The glacier reached southern Finland and Karelia in Russia (cf. Saarnisto & Lunkka 2004, Svendsen et al. 2004). During that glacial phase, the dark grey till and older drumlins were deposited in the area of southern Lapland. The marginal formations in Pudasjärvi and towards the northwest were deposited during the melting phase of that glaciation.

Following the melting phase, the climate became warmer. The interstadial began about 55–50 ka ago and lasted up to 30–25 ka ago, until the next glacial phase of the Late Weichselian began. The spread of the last glacier was rapid because the Weichselian maximum was reached during the next ten thousand years. Also, the melting phase was fast and shorter than the advance stage.

This interpretation means that radiocarbon ages presented in the literature (e.g. Korpela 1969, Hirvas 1991, Sutinen 1992) that cluster around 36–55 ka from organic inter-till layers are true. This idea is also supported by the TL ages clustering around 37 +/– 0.5 ka – 55 ka and OSL ages clustering around 41 +/– 2 ka – 66 +/– 5 ka, from which samples were taken from the same sections as the radiocarbon samples (Hütt et al. 1984, Mäkinen in press). This interpretation is also supported by age determinations of mammoth and reindeer bones (22.5–34 ka) (Ukkonen et al. 1999, Lunkka et al. 2001).

Discussion

Rapid warming of the climate during deglaciation probably caused the strong melting of the glacier's margin and surface. Particularly, the increase of mean temperature was almost 10°C during the few hundred of years after the Younger Dryas (12.8–11.6 ka ago) (cf. Boulton et al. 2001, Bard 2002). An imbalance between

the warm surface and cold bottom of the moving ice sheet followed and might be the reason for the initiation of ribbed moraine formation in the core areas of the latest glaciers in the northern hemisphere. Similar rapid climate change has not happened since then, so it could be an explanation why the ribbed moraines are not present in modern glaciated areas.

Acknowledgements

I am grateful to Prof. Matti Eronen, Prof. Matti Saarnisto and Dr. Keijo Nenonen for helpful comments on the manuscript and to Mrs. Carrie Turunen for checking the English.

REFERENCES

Aario, R. 1977. Classification and terminology of morainic landforms in Finland. Boreas 6, 87–100.

Bard, E. 2002. Abrupt climate changes over millennial time scales: climate shock. Physics Today 55, 32–38.

Boulton, G.S., Dongelmans, P., Punkari, M. & Broadgate, M. 2001. Palaeoglaciology of an ice sheet through a glacial cycle: the European ice sheet through the Weichselian. Quaternary Science Reviews 20, 591–625.

Helmens, K.F., Räsänen, M.E., Johansson, P.W., Jungner, H. & Korjonen, K. 2000. The Last Interglacial-Glacial cycle in NE Fennoscandia: a nearly continuous record from Sokli (Finnish Lapland). Quaternary Science Reviews 19, 1605–1623.

Hirvas, H. 1991. Pleistocene stratigraphy of Finnish Lapland. Geological Survey of Finland, Bulletin 354, 123 p.

Hütt, G., Punning, J.M, & Raukas, A. 1984. Application of the TL method to elaborate the geochronological scale of the late Pleistocene for the East European Plain. In: Mahaney, W.C. (ed.), Correlation of Quaternary chronologies. GeoBooks, United Kingdom, 47–55.

Hättestrand, C. 1997. Ribbed moraines in Sweden – distribution pattern and paleoglaciological implications. Sedimentary Geology 111, 41–56.

Kleman, J. 1994. Preservation of landforms under ice sheets and ice caps. Geomorphology 9, 19–32.

Korpela, K. 1969. Die Weichsel-Eiszeit und ihr Interstadial in Peräpohjola (nördliches Nordfinnland) im Licht von submoränen Sedimenten. Annales Academiae Scientiarum Fennicae A III: 99, 1–108.

Kujansuu, R. 1967. On the deglaciation of western Finnish Lapland. Bulletin de la Commission Géologique de Finlande 232, 98 p.

Kujansuu, R. 1975. Marrasjärven interstadiaalinen harju Keski-Lapissa (Summary: Interstadial esker at Marrasjärvi, Finnish Lapland). Geologi 27, 45–50. [in Finnish]

Lundqvist, J. 1969. Problems of the so-called Rogen moraine. Sveriges Geologiska Undersökningar C 648, 1–32.

Lundqvist, J. 1992. Glacial stratigraphy in Sweden. In: Kauranne, K. (ed.), Glacial stratigraphy, engineering geology and earth construction. Geological Survey of Finland, Special Paper 15, 43–59.

Lunkka, J.P., Saarnisto, M., Gey, V., Deminov, I. & Kiselova, V. 2001. Extent and age of the Last Glacial Maximum in the southeastern sector of the Scandinavian Ice Sheet. Global and Planetary Change 31, 407–425.

Lunkka, J.P., Johansson, P., Saarnisto, M. & Sallasmaa, O. 2004. Glaciation of Finland. In: Ehlers, J. & Gibbard, P.L. (Eds.), Quaternary Glaciations – Extent and Chronology. Amsterdam: Elsevier, 93–100.

Mäkinen, K. 1979. Interstadiaalinen turvekerrostuma Tervolan Kauvonkankaalla (Summary: An interstadial peat layer at Kauvonkangas, Finnish Lapland). Geologi 31, 82–87. [in Finnish]

Mäkinen, K. 1985. On the till-covered glaciofluvial formations in Finnish Lapland. Striae 22, 33–40.

Mäkinen, K. 1999. Ice wedge casts in Finnish Lapland. In: Seppälä, M. & Eerola, M. (eds.), Nordic symposium on changes in permafrost and periglacial environment : scientific and technical approach, Kevo, Finland, 2–24 August 1999 : programme, participants, abstracts, excursion guide. National Committee for Permafrost Research and Technics, 1 p.

Mäkinen, K. Dating the Weichselian deposits of the southwestern Finnish-Lapland. Geological Survey of Finland, Special Paper 40, in press (this publication).

Nenonen, K. 1995. Pleistocene stratigraphy and reference sections in southern and western Finland. Geological Survey of Finland, Miscellaneous publications, Ph.D. Thesis, 205 p.

Punning, J.-M. & Raukas, A. 1983. The age of tills: Problems and methods. In: Evenson, E.B., Schlüchter, Ch. & Rabassa, J. (eds.), Tills and related deposits. Rotterdam: Balkema, 357–364.

Saarnisto, M. & Lunkka, J.P. 2004. Climate variability during the last interglacial-glacial cycle in NW Eurasia. In: Battarbee, R.W., Gasse, F. & Stickley, C.E. (eds.), Past climate variability through Europe and Africa. Developments in Paleoenvironmental Research 6. Dordrecht: Kluwer Academic Publishers, 443–464.

Sarala, P. 2003. Ribbed-moreenit – jäätikön liikesuunnan poikittaiset indikaattorit (Summary: Ribbed moraines – transverse indicators of the ice flow direction). Geologi 55, 250–253. [in Finnish]

Sarala, P. 2005. Glacial morphology and dynamics with till geochemical exploration in the ribbed moraine area of Peräpohjola, Finnish Lappand. Geological survey of Finland, Espoo, 17 p.

Sarala, P., Peuraniemi, V. & Aario, R. 1998. Glacial geology and till geochemistry in ore exploration in the Tervola area, southern Finnish Lapland. Bulletin of the Geological Society of Finland 70, 19–41.

Sarala, P. & Rossi, S. 2000. The application of till geochemistry in exploration in the Rogen moraine area at Petäjävaara, northern Finland. Journal of Geochemical Exploration 68, 87–104.

Siivonen, L. 1975. New results on the history and taxonomy of the mountain, forest and domestic reindeer in Northern Europe. Biological Paper of University of Alaska, Special Reports 1, 33–41.

Svendsen, J., Alexanderson, H., Astakhov, V., Demidov, I., Dowdeswell, J., Funder, S., Gataullin, V., Henriksen, M., Hjort, C., Houmark-Nielsen, M., Hubberten, H., Ingolfsson, O., Jakobsson, M., Kjær, K., Larsen, E., Lokrantz, H., Lunkka, J.-P., Lyså, A., Mangerud, J., Matiouchkov, A., Murray, A., Möller, P., Niessen, F., Nikolskaya, O., Polyak, L., Saarnisto, M., Siegert, C., Siegert, M., Spielhagen, R. & Stein, R. 2004. Late Quaternary ice sheet history of northern Eurasia. Quaternary Science Reviews 23, 1229–1271.

Sutinen, R. 1992. Glacial deposits, their electrical properties and surveying by image interpretation and ground penetrating radar. Geological Survey of Finland, Bulletin 359, 123 p.

Ukkonen, P., Lunkka, J.P., Jungner, H. & Donner, J. 1999. New radiocarbon dates from Finnish mammoths indicating large ice-free areas in Fennoscandia during the Middle Weichselian. Journal of Quaternary Science 14, 711–714.

Quaternary studies in the northern and Arctic regions of Finland
Edited by Antti E.K. Ojala
Geological Survey of Finland, Special Paper 40, 67– 78, 2005.

DATING THE WEICHSELIAN DEPOSITS OF SOUTHWESTERN FINNISH LAPLAND

by
Kalevi Mäkinen

Geological Survey of Finland, P.O.Box 77, FIN-96101, Rovaniemi, FINLAND
E-mail: kalevi.makinen@gtk.fi

Mäkinen, Kalevi, 2005. Dating the Weichselian deposits of southwestern Finnish Lapland. *Geological Survey of Finland, Special Paper 40*, 67–78, 3 figures and 3 tables.

The Quaternary stratigraphy of Finnish Lapland is well defined where till deposits are concerned, but the comparative ages of the interstadial deposits between till beds have long been the subject of inconclusive research. Interstadials have been correlated with the Brörup and Odderade Interstadials of the southern edge of the Scandinavian ice sheet, which have been interpreted as corresponding to marine isotope stages MIS 5c and MIS 5a. Deposits at Sokli have been divided into three stages, of which the youngest has been correlated with the MIS 3 interstadial. The Sargejohka Interstadial of Finnmark, Norway, has also been correlated with MIS 3. A common feature among these deposits is signs of birch-dominated sub-arctic open tundra vegetation, exhibiting varying proportions of *Betula* and *Pinus* pollen. Of a total of 43 dating results obtained for these deposits using modern methods, only 9 exceed 59 ka in age. Results obtained using ^{14}C dating are located at the limit of the method's range of accuracy, so some results are finite, while others are infinite. These results may necessarily not be erroneous; rather they may indicate that the deposits were actually deposited towards the end of the Middle Weichselian about 45 – 59 ka ago. Datings obtained using TL and OSL methods were much younger than the expected ages corresponding to the Early Weichselian. Even after corrective coefficients were used in dating, the ages did not unambiguously coincide with the expected ages corresponding to MIS 5a and 5c.

The dating results obtained for the interstadial deposits of SW Lapland using modern methods do not support the hypothesis that the deposits were deposited more than 74 ka ago. Datings obtained using the ^{14}C, TL and OSL methods tend rather to indicate that the Peräpohjola-type interstadial deposits were formed during MIS 3. It has previously been assumed that till bed III was deposited during MIS 5d, but if the Peräpohjola Interstadial deposits of Finnish Lapland are interpreted as having been deposited during MIS 3, this would infer that till bed III may only have been deposited during MIS 4. This would indicate that Lapland was continually free of ice from the Eem Interglacial until the end of the Odderade Interstadial.

Key words (GeoRef Thesaurus, AGI): glacial geology, stratigraphy, till, organic sediments, interstadial environment, absolute age, relative age, Pleistocene, Weichselian, Lapland Province, Finland

Introduction

Finnish Lapland lies in the central part of the Fennoscandian glaciations, and has been completely covered by ice several times over the last couple of million years. Six distinct till beds (I–VI) deposited during different flow stages have been observed in Lapland (Hirvas et al. 1977). These till beds differ

both in their overall structure, and in the orientation of stones indicating the direction of ice flow. Very little is known about the earliest of these glaciations, but many more observations have been made of the till deposits and the interlaying silty, sandy, and organic deposits deposited during the Weichselian Glaciation, (Hirvas 1991, Hirvas et al. 1976, Mäkinen 1981, 1985, Sutinen 1982, 1984, Kujansuu 1975, Kujansuu & Eriksson 1995, Ilvonen 1973, Helmens et al. 2000).

The Weichselian Glaciation in Lapland can be divided into three flow stages, with the latter two stages closely connected (Hirvas et al. 1977). During the Middle Weichselian and the Late Weichselian, flow stage II produced till bed II, which is widespread throughout Finnish Lapland. Flow stage I is connected to the deglaciation of flow stage II, which culminated in the formation of till bed I, deposited during local oscillations and by local ice lobes. Till bed I is mainly found in northern and eastern parts of Finnish Lapland (Hirvas 1991). Till bed III was deposited in the Early Weichselian during flow stage III (Hirvas 1991). Till beds II and III are widely distributed over southwestern Lapland (Korpela 1969, Mäkinen 1981).

The geomorphological landforms of S and SW Lapland were partly formed during flow stage III and flow stage II. During flow stage III ice flowed from NNW, or N in places (Hirvas et al. 1976, Korpela 1969, Mäkinen 1979), as far as Pudasjärvi, in Ostrobothnia, where extensive marginal formations mark its southernmost extent (Sutinen 1982, 1984). Drumlin fields at Kemi and Ranua were formed during this flow stage (III), with the drumlins aligned NNW-SSE. During the deglaciation phase of the same flow stage, eskers aligned NW-SE were formed in SW Lapland, but these eskers were subsequently covered with till during the flow stage II in the Middle and Late Weichselian (Mäkinen 1985).

During the Weichselian Glaciation, western and southern Lapland were free of ice at least once, whereas eastern and northern Lapland may have been ice-free twice (Hirvas 1991). Deposits surveyed at Sokli indicate that the ice may have disappeared from certain eastern areas during as many as three phases of the Weichselian Glaciation (Helmens et al. 2000). Two separately deposited series of organic deposits have also been found in Swedish Lapland and in Finnmark, Norway, indicating that these regions were free of ice twice during the Weichselian (Lagerbäck 1988, Lagerbäck & Robertsson 1988, Hirvas et al. 1986).

The first organic deposits to be found beneath the till in Lapland were gyttja deposits at Permantokoski (Fig. 1), as described by Korpela (1969). Their pollen flora is dominated by *Betula* (birch), indicating the presence of sub-arctic vegetation. Korpela (1969) gave the name Peräpohjola Interstadial to the ice-free period when these gyttja deposits were deposited, and compared it to the contemporary Jämtland Interstadial defined by Lundqvist (1967) in Sweden. Lundqvist has correlated the Jämtland Interstadial to the Brörup Interstadial described by Andersen (1961) in Denmark. Many more interstadial organic deposits of the Weichselian age have subsequently been found in Lapland and northern Ostrobothnia (Kujansuu 1975, Hirvas et al. 1977, Mäkinen 1979, 1985, Sutinen 1982, 1984, Forsström 1988, Iisalo 1992, Nenonen 1995, Helmens et al. 2000). The pollen flora of these deposits varies greatly, with some rich in birch, others dominated by pine, and some containing plenty of pollen from shrubs and grasses.

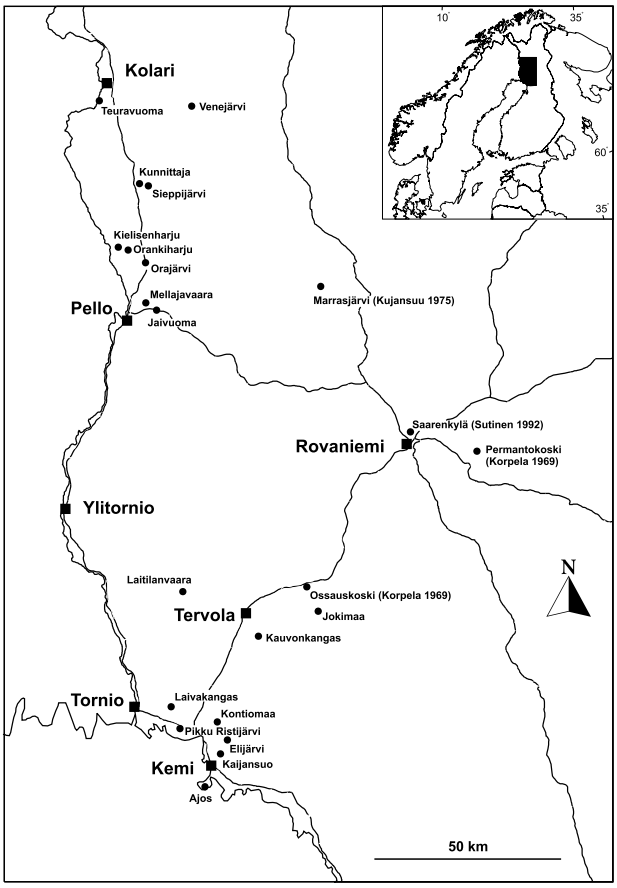


Fig. 1. Localities of dated minerogenic and organic deposits in SW Lapland (Map © National Land Survey of Finland, publication licence number: 529/MYY/05).

On the basis of their pollen flora composition, organic deposits can be classified as interglacial deposits containing pollen from birch, pine, alder, larch and broad-leaved trees; interstadial deposits with birch and pine pollen; or interstadial deposits more dominated by birch pollen. Interglacial deposits were mainly deposited during the Eem Interglacial that preceded the Weichselian Glaciation (e.g. Tepsankumpu Inter-

glacial, Hirvas 1991 and Saarenkylä Interglacial, Sutinen 1992); although at least one set of deposits was formed during the earlier Holstein Interglacial stage (Naakenavaara Interglacial, Hirvas 1991). These deposits are stratigraphically located beneath till bed III. The interstadial organic deposits lie between till bed II and till bed III (Hirvas et al. 1977, Mäkinen 1979, Sutinen 1984).

Dating methods and results

Until the 1960s the main dating method used in such analyses focussed on the composition of pollen and siliceous algae, and the locations of deposits in relation to stratigraphic till units. Since the 1960s ^{14}C dating method has been more widely used to date organic deposits. The thermoluminescence (TL) and opticially stimulated luminescence methods (OSL) have also been applied, particularly to date minerogenic deposits. ^{14}C dating techniques have meanwhile been improved more recently, and accelerator mass spectrometry (AMS) can today be used for material up to 75 ka old. Peat samples have also been dated using the uranium/thorium method (U/Th).

Pollen analyses

Both organic and sorted deposits can be found in SW Lapland beneath till bed II, which was deposited during the flow stage II (Korpela 1969, Mäkinen 1985, Sutinen 1984). The composition of the pollen flora within these organic deposits varies, but nevertheless generally resembles that of the Peräpohjola Interstadial deposits from Permantokoski and Ossauskoski (Korpela 1969), where birch accounts for 83–99% of tree pollen. In the pollen flora of the deposits from Kauvonkangas in Tervola (Mäkinen 1979, Fig. 5), birch accounts for 85 – 95% of tree pollen and some 55–75% of all pollen. In a new more detailed analysis of four samples from peat deposits, concentrations of *Betula* tree pollen and *Betula nana* pollen were assessed separately. The total proportions of *Betula* tree pollen were 0 – 35%, while *Betula nana* accounted for 8 – 31% of all the pollen (Fig. 2). In the lowest parts of the peat, all of the birch pollen consisted of *Betula nana*. In the upper layers of the deposits, the proportion of birch tree pollen rose to 35%. The proportion of *Betula nana* pollen was highest in the middle of the peat deposits, and declined to 8% higher up. The pollen flora also contained notable amounts of *Cyperaceae* herbs (38 – 60%) and *Salix* willows (12 – 13%).

^{14}C dating

In this research project a total of 20 samples of organic deposits laid down in SW Lapland during the Weichselian Glaciation have been dated using the ^{14}C method, producing a wide range of results (Table 1). The ages of dated samples of organic deposits from beneath till bed II vary in finite age from 42,800 – 52,600 years BP, and in infinite age from >45,000 – >55,000 years BP. The drawback of this method is that the maximum age for which this dating method is reliable is only about 40,000 – 50,000 BP years. Finite ages were only obtained for nine of the datings. Eleven of the dating results were so close to the background

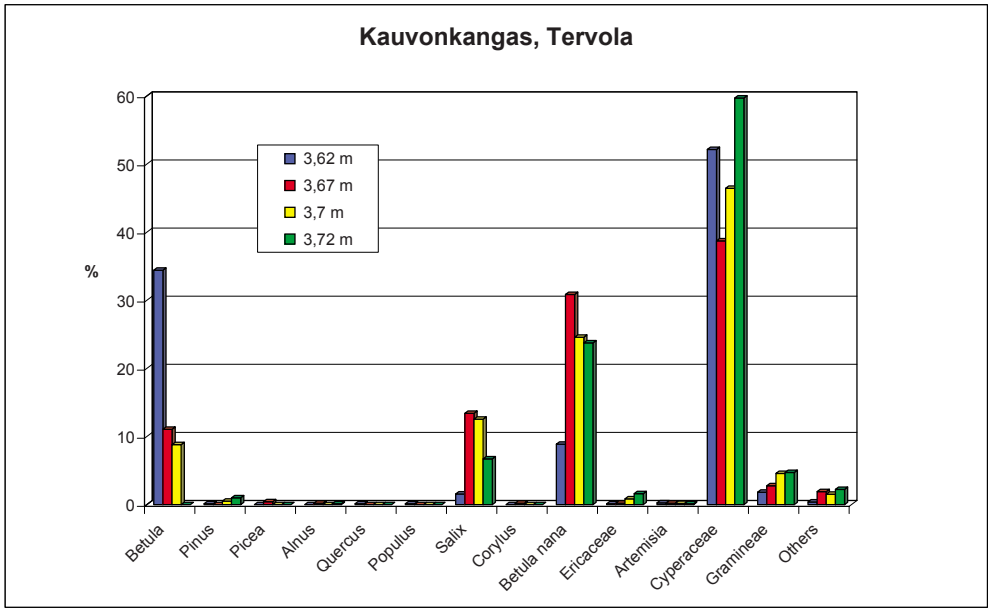


Fig. 2. Pollen analyses from Kauvonkangas Peat (Analysed: B. Eriksson).

Table 1. ^{14}C ages from SW-Lapland (A = humin fraction and B = humus fraction).

Locality	Depth (m)	Laboratory number	Age years BP
Kauvonkangas	3,7	Su-657	>49,000
Kauvonkangas	3,55	Su-688	48,000 $^{+4,100}_{-2,400}$
Kauvonkangas	3,8	Su-689	48,100 $^{+3,500}_{-2,400}$
Orankiharju	3,20	Su-752A	40,500 $^{+2,900}_{-2,200}$ (humiini)
	3,20	Su-752B	45,000 $^{+4,500}_{-3,000}$
Orajärvi I	4,10	Su-753A	49,500 $^{+4,000}_{-2,700}$
	4,10	Su-753B	>45,800
Orajärvi II	2,50	Su-754A	>51,700
	2,50	Su-754B	>51,300
Orajärvi III	5,0	Su-1091A	20,900 ± 200
	5,0	Su-1091B1	>46,800
	5,0	Su-1091B2	39,300 $^{+3,400}_{-2,400}$
Kunnittaja	3,0	Su-1185A	>52,000
	3,0	Su-1185B	>45,000
Orajärvi IV	5,0	Su-1186A	>45,000
	5,0	Su-1186B	42,900 ± 1.600
Ajos I	5,50	Su-1187A	40,600 ± 1100 (humiini)
Ajos I	7,0	Su-1188A	38,700 ± 900 (humiini)
		Su-1188B	46,000 $^{+2,900}_{-2,100}$ (humus)
Kunnittaja		Su-1280	51,600
Orajärvi V		Su-1282	>55,000
Ajos II	1,90	Su-1283	>53,000
Ajos III	2,70	Su-1284	42,800 ±1100
Kielisenharju I	4,0	Su-1782	>54,300
Kielisenharju II	3,20	Su-1797	>49,200
Jokimaa	3,0	Su-2234	51,500 ±3000
Jaivuoma	4,70	Su-2454	>54,900
Jaivuoma	7,85	Su-2455	52,600 ±2300
Permantokoski		I-1139	>28,000 (Korpela 1969)
Permantokoski		K-722	>35,000 (Korpela 1969)
Permantokoski		GrN-4543	>42,700 (Korpela 1969)
Ossauskoski		I-1138	>30,000 (Korpela 1969)
Kostonniska, Taivalkoski		GrN-4491	45,400 ±2000 (Korpela 1969)
Marrasjärvi		Su-236	42,300 $^{+6,000}_{-1,700}$ (Heikkinen 1975)
Marrasjärvi		Su-237	48,000 $^{+4,700}_{-2,300}$ (Heikkinen 1975)
Marrasjärvi		Su-263	>55,000 (Heikkinen 1975)

values for the method that only infinite ages could be ascertained (e.g. > 45,000 years BP).

Uranium/thorium dating

Vogel & Kronfeld (1980) introduced a new method for dating peat, based on the preferential uptake of uranium (U) by peat due to its high solubility in groundwater relative to thorium (Th). This method has subsequently been used and developed by researchers including Van der Wijk et al. (1986), Van der Wijk et al. (1988), Heijnis (1992) and Heijnis & Van der Plicht (1992). The U/Th disequilibrium method is based on disequilibrium between ^{234}U and its product ^{230}Th. This allows the dating of samples up to 350 ka old (Heijnis 1992). The most reliable results can be obtained for samples taken from deposits formed as closed geochemical systems where thorium concentrations were

zero, or at a known level, when the peat was formed (Ivanovich & Harmon 1982, Heijnis 1992, Heijnis & Van der Plicht 1992).

U/Th dating results for samples from peat deposits at Kauvonkangas in Tervola have been published in several articles. Van der Wijk et al. (1986) dated two peat samples from Tervola, of which one was dated at 45 ± 5,1 ka old. The chemical properties of these two samples differed so much from each other that they could not prove that the peat samples from Tervola had come from a closed system. This means that the dating results are more subject to chance. Van der Wijk (1987) interpreted the sampling environment as open geochemical system where samples could not be reliably dated. Heijnis (1992) has also examined opportunities to date peat samples from open geochemical systems using special corrective coefficients. It has not been possible to date the peat deposits at Tervola accurately, but the results indicate they are 80–100 ka old (Heijnis 1992). Three further peat samples from Tervola have been dated (in the Laboratory of Niedersächsisches Landesamt fuer Bodenforschung C-14 and H-3, Geyh personal communication 1990), giving age results of 92,7 ± 3,4, 95,0 ±5,2 and 93,0 ± 3,5 ka old, with a corrected average value of 93,0 ± 2,2 ka old.

Thermoluminescence and optically stimulated luminescence dating

Luminescence dating methods are based on measures of the radiation doses contained in quartz and feldspar crystals in sediments. In the natural environment, these mineral crystals are able to absorb and accumulate radiation energy when they are exposed to alpha, beta and gamma radiation from radioactive materials in their surroundings. When these minerals are exposed to light, the stored energy will be released and emitted as a blue-violet signal called luminescence. The signal is called TL if the energy is released by heat (500°C) and OSL if the release is by light (Olsen et al. 1996). When the minerals are exposed to sunlight during the transport or sedimentation in nature the energy is released and the latent signal is reduced to zero (or almost to zero). When the zeroed minerals are buried deeper inside deposits, their continuous exposure to radiation in their surroundings again increases their energy. If the intensity of the irradiation is constant, the radiation dose absorbed (called paleodose) will be proportional to the time elapsed since the zeroing even took place. This time period (the age) is given by the following equation: Age = Paleodose/Annual dose (Olsen et al. 1996).

TL datings have been conducted at the Nordic Laboratory for Luminescence Dating, in Risö, Denmark (Mejdahl 1991), and at the Dating Laboratory of the Academy of Sciences in Tallinn, Estonia (Hütt et al. 1982). The TL dating results are compiled in Table 2. The ages of the samples dated at Risö were calculated at between 37 ± 5 ka and 55 ± 5 ka. The ages of the samples dated in Tallinn varied from 55 to 120 ka old. OSL analyses were also conducted in Tallinn, with the results shown in Table 3. The youngest sample, dated at 20 ± 9 ka old, came from an ice-marginal formation at Mellajavaara in Pello; while the oldest sample, dated at 117 ± 15 ka old, were from a till-covered glaciofluvial

Table 2. TL ages from SW-Lapland (R=Risø, T=Tallinn).

Locality	Depth (m)	Laboratory number	TL-age (ka)	Corr. age (ka)
Kontiomaa	1,7	R-893203	38 ± 5	52 ± 7
Ajos A	3,5	R-893204	49 ± 5	67 ± 7
Ajos B	2,0	R-893205	46 ± 5	63 ± 7
Ajos B	3,5	R-893206	52 ± 5	71 ± 7
Orajärvi A	1,9	R-893207	45 ± 5	61 ± 7
Orajärvi B	2,3	R-893208	50 ± 5	68 ± 7
Kielisenharju	2,5	R-893209	44 ± 5	60 ± 7
Kielisenharju	2,8	R-893210	55 ± 5	75 ± 7
Kauvonkangas	5,5	T-	55	75
Kauvonkangas	7,4	T-	74	102
Kauvonkangas	12,0	T-	120	163

R- (Mejdahl 1991, Olsen et al. 1996)
T- (Hütt et al.1982)

Table 3. OSL ages from SW-Lapland.

Locality	Depth (m)	Laboratory Number	OSL-age (ka)	Corr. age (ka)
Mellajavaara	1,5	91040	20 ± 9	28
Kaijansuo	2,7	91393	41 ± 2	56
Teuravuoma	2,5	91407	43 ± 6	59
Venejärvi	2,2	91408	68 ± 9	93
Sieppijärvi	1,8	91412	69 ± 9	94
Jokimaa	2,3	91433	104 ± 13	141
Laitilanvaara	3,0	91437	106 ± 11	144
Laivakangas	1,5	91450	117 ± 15	159
Pikku Ristijärvi	1,5	91451	45 ± 6	62
Elijärvi	2,1	91453	86 ± 6	117
Kauvonkangas	3,0	91434-1	66 ± 5	90
Kauvonkangas	2,8	91434-2	–	
Kauvonkangas	2,5	91434-3	57 ± 5	78

Sample 91434-2 was poor for feldspar's grains = 100 microns

marginal formation at Laivakangas in Tornio. Most of the dating results fell within the age range 41 – 69 ka. A significant exception was a set of datings from Elijärvi, Laitilanvaara, Jokimaa and Laivakangas, which fell within the age range 86 – 117 ka.

Discussion

It is widely believed that during glaciation in the central and eastern parts of Fennoscandia there were one or two interstadials, linked to the Early Weichselian Brörup Interstadial (MIS 5c) and/or the Odderade Interstadial (MIS 5a) (Hirvas et al. 1986, Lagerbäck & Robertsson 1988, Nenonen 1995, Saarnisto & Lunkka 2004, Svendsen et al. 2004). In northern Sweden, two interstadials have been described, of which the older correlates with the Brörup Interstadial, while the younger Tärendö Interstadial has been correlated with the Odderade Interstadial (Lagerbäck & Robertsson 1988). Earlier research in Finnish Lapland has largely focussed on a single interstadial (Hirvas et al. 1977, Sutinen 1984, Mäkinen 1985, Hirvas 1991), generally thought to correspond to the Brörup Interstadial (Donner et al. 1979). Nenonen (1995) has correlated Peräpohjola with Odderade Interstadial. Helmens et al. (2000) have nevertheless described a series of deposits at Sokli in eastern Lapland indicating the occurrence of three interstadials. The most recent of these is believed to have occurred during MIS 3. It is thought that till beds III and II from the Weichselian Glaciation in Lapland were deposited during MIS 5d and MIS 4-MIS 2, respectively.

The vegetation during the Peräpohjola Interstadial has so far been described as birch tree dominated (*Betula* 85–99 %), since earlier analyses focussed mainly on tree pollens, with pollens of all birch varieties, including *Betula nana* (dwarf birch), calculated together as birch tree pollen. The composition of the pollen flora varies, but nevertheless generally resembles that of the deposits from Permantokoski and Ossauskoski (Korpela 1969), where birch accounts for 85–99% of tree pollen and some 50–80% of all pollen. In the pollen flora of the deposits from Kauvonkangas in Tervola (Mäkinen 1979), birch accounts for 85–95% of tree pollen and some 55–75% of all pollen. In a later more detailed analysis of four samples from Kauvonkangas peat deposits, concentrations of *Betula* tree pollen and *Betula nana* pollen were assessed separately. The total proportions of *Betula* tree pollen were 0 – 35% of all pollen, while *Betula nana* accounted for 8 – 31% of all the pollen (Fig. 2). In the lowest parts of the peat, all of the birch pollen consisted of *Betula nana*. In the upper layers of the deposits, the proportion of birch tree pollen rose to 35%. The proportion of *Betula nana* pollen was highest in the middle of the peat deposits, and declined to 8% higher up. The pollen flora also contained notable amounts of *Cyperaceae* herbs (38

– 60%) and *Salix* willows (12 – 13%). Pollen flora analyses of the Kauvonkangas Interstadial, a type locality for Peräpohjola Interstadial, indicate the presence of willow stands and dwarf birch heaths, as well as watery, sedge-dominated mires and grasslands (Eriksson 2005). Helmens et al. (2000) have described interstadial deposit (MIS 3) from Sokli, and its pollen flora corresponds to shrub-tundra vegetation, with *Betula* and *Betula nana* (c. 10%) accounting together for 35% of all the pollen, *Pinus* 10%, *Salix* 5% and *Ericales* 30–40%. Pollen flora of a similar composition has been found in organic deposits overlying till-covered marginal formations at Ajos. The pollen in the lower parts of Ajos peat deposits is 40 – 60% *Betula* (including *Betula nana*), >10% *Pinus*, 15 – 38% Ericaceae and 15–20 % Cyperaceae of all pollen. In the upper parts the proportion of *Betula* pollen rises to about 80% (including *Betula nana*). Pollen flora compositions in most of the organic deposits from SW Lapland correlate with those of Permantokoski and Ossauskoski found by Korpela (1969). Analysis of the pollen flora from Maaselkä Interstadial (Hirvas 1991) indicates the presence of very similar vegetation, even though *Betula nana* pollen was not analysed separately from other types of birch pollen (Eriksson 2005).

The pollen flora profiles of the interstadial deposits in Lapland and Ostrobothnia differ considerably. The pollen flora of the interstadial deposits in Ostrobothnia is mainly dominated by *Pinus*, whereas in Lapland it is dominated by *Betula*. Forsström (1988) has discussed these differences, and correlated the *Pinus*-dominated interstadial deposits in Oulainen and Ruottisenharju with the Brörup Interstadial; and the *Betula*- dominated deposits in Permantokoski (Korpela 1969), Kostonniska (Korpela 1969) and Kauvonkangas (Mäkinen 1979) with the Odderade Interstadial. Nenonen (1995) has correlated both the Peräpohjola Interstadial and the *Betula*-dominated Mertuanoja Interstadial in Ostrobothnia with the Odderade Interstadial.

Researchers examining Lapland's interstadial deposits have repeatedly encountered doubts about how many separate interstadial stages occurred in Lapland during the Weichselian glaciation, and about which marine isotope stages these interstadials correspond to. Where dating is concerned, researchers have long depended on ^{14}C dating, and many deposits of infinite age assessed on the basis of their pollen flora composition have been assigned to the Early Weichselian interstadials described for the southern fringes of the glaciated regions of Scandinavia (Donner 1983, Hirvas et al. 1986, Forsström 1988, Hirvas 1991, Nenonen 1995). Only since the late 1980s have interstadial deposits been dated using new methods, in some cases giving results that have differed considerably from their expected ages.

When dating results obtained for deposits from SW Lapland using the ^{14}C, TL and OSL methods are compared with the deep-sea oxygen isotope stages (Fig. 3), it becomes evident that their ages correspond with MIS 3, with ages ranging 37 – 57 ka, with the excep-

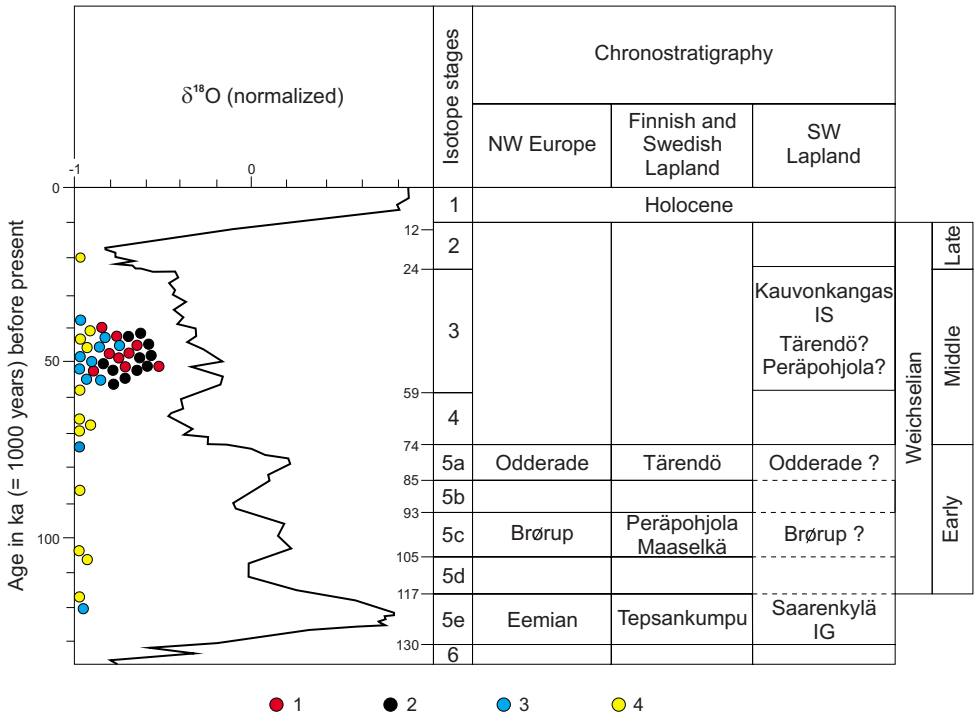


Fig. 3. Ages obtained through ^{14}C, TL and OSL dating from SW Lapland compared with the deep sea isotope curve (Martinson et al. 1987). Symbols 1 = finite ^{14}C age, 2 = infinite ^{14}C age, 3 = TL age and 4 = OSL age. Chronostratigraphy in Finnish and Swedish Lapland after Hirvas (1991) and Lagerbäck & Robertsson (1988).

tion of eight OSL datings and two TL datings, which produced results of 20 ka (MIS 2) and of 66 – 68 ka (MIS 4), 74 ka (MIS 5a), 86 ka (MIS 5b), 104 ka (MIS 5c), 106 ka (MIS 5d), 117 – 120 ka (MIS 5e).

Till-covered glaciofluvial formations are common in SW Lapland (Mäkinen 1985), and organic deposits whose pollen flora is dominated by *Betula* have been found both in the lower parts of till bed II, and between till bed II and the underlying glaciofluvial formation. These organic deposits have been dated by the ^{14}C method. All of the nine finite ages (from 42,800 – 52,600 years BP, Table 1) corresponded to the deep-sea oxygen isotope stage MIS 3. The eleven infinite datings (from >45,000 – >55,000 years BP) indicate the minimum ages of samples, but cannot reveal whether samples are for instance 45 – 60 ka old or 74 – 105 ka old. Since these interstadial deposits have widely been assumed to correlate with the Brörup Interstadial or possibly the Odderade Interstadial on the southern edge of the ice sheet, they have been assumed to be at least 74 ka old, and at most 105 ka old. The ^{14}C dating results from the same till-covered glaciofluvial ice-marginal formation vary between finite and infinite ages. For instance, the ages of the gyttja deposits at Ajos range from 42,800 ±1,100 (Su-1284) to >53,000 (Su-1283) years BP (Mäkinen 1985). At Kauvonkangas, the interstadial peat layer was initially dated at >49,000 years BP (Su-657) and later new samples from the same peat layer gave ages of 48,000 $^{+4,100}_{-2,400}$ (Su-688) and 48,100 $^{+3,500}_{-2,400}$ years BP (Su-689, Table 1). Five datings have been conducted on the gyttja deposits at Orajärvi, all of which gave infinite ages from >45,000 to >55,000 years BP. Only the gyttja deposits at Orankiharju (40,500 $^{+2,900}_{-2,200}$ (Su-752A), 45,000 $^{+4,500}_{-3,000}$ (Su-752B) years BP) and Jokimaa (51,500 ±3,000 (Su-2234) years BP) gave finite ages, which are very near the background values of the ^{14}C dating method. Both Kujansuu (1975) and Hirvas (1991) have dated Weichselian organic deposits from Lapland. The ^{14}C dating gave an age of >55,000 years BP (Su-263) for organic deposits found in a till-covered interstadial esker at Marrasjärvi (Kujansuu 1975, Heikkinen 1975). Hirvas (1991) has published 8 radiocarbon dates from interstadial-type deposits from Lapland, only one of which is infinite – >49,500 years BP (Su-874). The other datings vary between 41,900 ±3,000 years BP (Su-508) and 49,000 $^{+4,600}_{-3,000}$ years BP (Su-725). ^{14}C datings have, however, been shown to be unreliable for samples more than 40,000 years (BP) old (Mangerud 1991, Donner 1995). In 40,000-year-old samples, the amounts of radiation that can be measured are so small that slight fluctuations can lead to great variations in age estimates. Moreover, just a small amount of younger organic matter (e.g. plant roots) in a sample will result in its dating being considerably younger than its real age (Kankainen & Huhta 1986). The large number of infinite ages obtained through the ^{14}C method do not necessarily indicate that the dated interstadial samples are considerably older than the ages calculated; rather they show that the samples' ages are either very near the maximum age that can be verified using this method, or older. Due to these factors, it seems evident that the ^{14}C method alone cannot give reliable ages for the Peräpohjola interstadial deposits. Therefore other dating methods should also be used to produce more accurate dating results.

The ages of the peat samples from Kauvonkangas obtained using the U/Th method are fairly random, since the samples came from an open geochemical system. The peat deposits at Kauvonkangas are covered by thin layers sandy littoral deposits and sandy till, through which water can seep into the peat. Although these peat deposits have been dated at 92 ka old (Geyh 1990), with corrective coefficients allowing for an age range of 80 – 100 years (Heijnis 1992), this author would prefer to disregard these results, since they are too unreliable.

Ages obtained through TL dating indicate that the luminescence signals in the dated deposits were last reduced to zero about 38 – 55 ka ago, during MIS 3, with the exception of the silts (74 ka) and sands (120 ka) from beneath the Kauvonkangas peat deposits. In both the TL and OSL methods it is vital to take samples from deposits that were adequately exposed to sunlight when they were previously deposited, thus ensuring that their luminescence signals were completely reduced to zero. If this did not occur, the dating results will indicate that the deposits are considerably older than their true age (Olsen et al. 1996). Since the sandy deposits beneath the peat deposits at Kauvonkangas are of glaciofluvial origin, their luminescence signal may not have been completely reduced to zero before they were last redeposited, and this would explain why their dating results indicate they are dated as being so much older than samples taken from other deposits in the same place in the stratigraphic sequence.

The ages obtained using OSL dating have a wider range than those calculated using the TL method. Four samples were dated as corresponding to MIS 3, three to MIS 4, and one sample for each of MIS 2, MIS 5b, 5c, 5d and 5e. The youngest sample (20 ka) is thought to have been deposited during the Late Weichselian deglaciation stage, although its OSL signal had probably not been completely reduced to zero, so the age obtained probably pre-dated the deglaciation. The samples dated as corresponding to MIS 5 may similarly not have been sufficiently exposed to sunlight to

remove all their older radiation and reduce their OSL signals down to zero. Their calculated ages would indicate that they are from a stage that considerably predated their expected ages, and these ages do not always fit in with the stratigraphic location of their formations.

Since the ages obtained through the TL and OSL dating methods for the interstadial deposits thought to have been deposited during the Early Weichselian have been considerably younger than their deep-sea isotope stage ages, reasons for this discrepancy have been sought. One possible reason may be the effect on the dating of shallow traps in the surfaces of the mineral structures being dated. Mejdahl et al. (1992) have shown that dating results should be corrected for the shallow trap effect according to the formula $Y = 1.35X + 0.70$, where X is the uncorrected age, and Y the corrected age. Corrected TL ages (Table 2) for deposits from SW Lapland have been calculated using this formula and subsequently published by Olsen et al. (1996). They point out that the need for age corrections is greatest for older samples (ages above 80 – 100 ka). It is uncertain whether significant underestimation of ages normally occurs for low ages (Olsen et al. 1996).

The corrected TL and OSL dating results for SW Lapland correlate their deposits with MIS 3 (for 2 samples), MIS 4 (for 8 samples), MIS 5a (for 3 samples), MIS 5b (for 2 samples), MIS 5c (for 2 samples), MIS 5e (for 1 sample) and MIS 6 (for 3 samples). The greatest readjustments in the ages obtained concern the reallocation of 8 datings to stadial MIS 4. Since all the dated samples have come from stratigraphic locations between till beds II and III, the corrected results are still inconsistent. Although Olsen et al. (1996) have corrected the dating results for fading, they suspect that their correction coefficient may have been too large. This hypothesis is backed up by Visocekas's (2000) research, which indicates that luminescence signals in Finnish feldspars fade less rapidly than has been assumed.

As the interstadial deposits of SW Lapland are compared to either the Brörup Interstadial or the Odderade Interstadial, their expected ages should be at least 74 ka old. Of the 23 dated samples from SW Lapland, 17 uncorrected TL or OSL datings are younger than the minimum expected age for the Odderade Interstadial; one result places a sample at the lower boundary of MIS 5a; one corresponds to MIS 5c; and one to MIS 5d. After correcting all the TL and OSL dating results using the formula described above, eleven of the dating results still fall below the minimum age for the Odderade Interstadial , and only three of the results that exceed this age are placed in the age range corresponding to the Odderade and Brörup Interstadials. Thirteen uncorrected results and three corrected results correspond to MIS 3.

Dating results have been published for 24 OSL samples, 23 TL samples and 1 U/Th sample from Ostrobothnia (Jungner 1987, Donner 1988, Aalto et al. 1989, Gibbard et al. 1989, Niemelä & Jungner 1991, Kujansuu et al. 1991, Hütt et al. 1993, Nenonen 1995). All the ages, with the exception of one obtained through TL dating, indicate that the luminescence signal of the dated sediments were last reduced to zero during the Late Saalian (MIS 6) and the Eemian (MIS 5e). The divergent sample was deposited during the Brörup interstadial (MIS 5c). The ages obtained using OSL dating have a wider range than those calculated using the TL dating method. Six samples were dated as corresponding to MIS 6 and MIS 5e; ten to MIS 5c; one to MIS 4; and three to MIS 3. A comparison between the results obtained using the two dating methods shows that the OSL dates are generally younger than the TL dates for the same samples (Hütt et al. 1993).

There are clearly great differences between the TL and OSL dating results from SW Lapland and Ostrobothnia. This difference cannot be the consequence of methodological differences, and can only be explained as being an indication of a real age difference between the dated samples. In Ostrobothnia, the signals in most of the dated deposits were last reduced to zero during MIS 6, MIS 5e and MIS 5c; while for most of the dated samples in Lapland this did not occur earlier than during MIS 3. This discrepancy may be explained by the fact that Ostrobothnia was free of ice from the deglaciation of the Late Saalian Glaciation to the end of MIS 5c (Nenonen 1995), or possibly even to the end of MIS 5a. SW Lapland was covered by ice at least once during the Early or Middle Weichselian, and the dated samples from this region were deposited after this stadial, probably during MIS 3 or MIS 5a. This could mean that SW Lapland was free of ice from the deglaciation of the Late Saalian Glaciation to the end of MIS 5a, and also after the MIS 4 stadial, during MIS 3.

Helmens et al. (2000) have shown through the ^{14}C -AMS dating of wood fragment that the region around Sokli in eastern Lapland was free of ice during MIS 3. Samples from southern and western Finland have been dated using various methods as originating from MIS 3. Deposits from the Mertuanoja Interstadial (Iisalo 1992, Nenonen 1995) have also been dated, with ^{14}C dating indicating that the humic fraction of the peat is 36,600 ± 700 years BP old. The OSL-dated age of the fine sand found in the Mertuanoja Peat was 64 ± 6 ka, and the sandy deposits covering the peat were dated as being 38 ± 6 ka old. According to Ukkonen

et al. (1999) dated mammoth bones from Iijoki, Haapajärvi, Nilsiä and Helsinki and Espoo (ages from ca. 32,000 to 22,500 years BP) and a reindeer antler from Tornio, in northern Finland (34,300 years BP; Siivonen 1975), both show that there was probably a large ice-free area in the SE part of Fennoscandia during the Middle Weicheslian substage (Ukkonen, et al. 1999). The age of layer IIa in Susiluola Cave, Kristiinankaupunki, calculated through infrared optically stimulated luminescence (IRSL) at approx. 35 ka, also indicates the occurrence of a corresponding ice-free period during MIS 3 (Schulz et al. 2002). The age of the Sargejohka Interstadial in Finnmark (Olsen et al. 1996) has been calculated through ^{14}C -AMS dating as about 35,000 years BP, and through TL dating as 37 and 41 ka; and these ages also support the concept of a Middle Weichselian Interstadial.

According to Nenonen (1995) and Donner (1995), western and southern Finland was free of ice during the Early Weichselian, and the same areas were covered by an ice sheet during the Middle and Late Weichselian. According to Hirvas (1991), Lapland was glaciated during the Early Weichselian (MIS 5d), when the ice sheet spread as far as Pudasjärvi, in Ostrobothnia (Sutinen 1992). When Peräpohjola Interstadial and its type locality Kauvonkangas is correlated with MIS 3 so it opens the question, was till bed III deposited during the MIS 5d as proposed by Hirvas (1991) or during MIS 4 as proposed by Sarala (2005). If deposition took place during MIS 4 so it means that in SW Lapland, deposits from Eemian and possible subsequent Early Weichselian interstadials would situate between the tills of the Saale and Middle Weichselian (MIS 4).

Conclusion

It is not possible to define more than one series of interstadial deposits (Peräpohjola) in SW Lapland on the basis of their position in the stratigraphic sequence and their pollen flora. Pollen flora analyses of the Kauvonkangas Interstadial, a type locality for Peräpohjola Interstadial, indicate subarctic open tundra vegetation with the presence of willow stands and dwarf birch heaths, as well as watery, sedge-dominated mires and grasslands. Dating results obtained using various methods indicate that the dated samples from Peräpohjola Interstadial were deposited during MIS 3. Till bed II, which overlays the dated organic and sorted deposits, was deposited during the Late Weichselian (MIS 2). Till bed III, which lies beneath these deposits, is older than MIS 3, and was deposited during either the Middle (MIS 4) or the Early Weichselian (MIS 5d). If till bed III was deposited during the Middle Weichselian, this would mean that the Eem Interglacial and the possible subsequent interstadials of the Early Weichselian form an uninterrupted continuum, during which SW Lapland remained free of ice until the beginning of MIS 4.

Acknowledgements

The TL dating work for this study was funded by the Finnish Academy of Sciences and OSL dating work was financed by the Geological Survey of Finland. The radiocarbon samples were analysed by Tuovi Kankainen at the Radiocarbon Laboratory of the Geological Survey of Finland (Su-).

REFERENCES

Aalto, M., Donner, J., Hirvas, H. & Niemelä, J. 1989. An interglacial beaver dam deposit at Vimpeli, Ostrobothnia, Finland. Geological Survey of Finland, Bulletin 348, 34 p.

Andersen, S.Th. 1961. Vegetation and its Environment in Denmark in the Early Weichselian Glacial (Last Glacial). Danmarks geologiske undersøgelse II. Række 75, 175 p.

Donner, J. 1983. The identification of Eemian interglacial and Weichselian interstadial deposits in Finland. Annales Annales Academiae Scientiarum Fennicae, Series A III, Geologica – Geographica 136, 38 p.

Donner, J. 1988. The Eemian site of Norinkylä compared with other interglacial and interstadial sites in Ostrobothnia, western Finland. Annales Academiae Scientiarum Fennicae, Series A III, Geologica – Geographica 149, 31 p.

Donner, J. 1995. The Quaternary history of Scandinavia. World and Regional Geology 7. Cambridge: Cambridge University Press, 200 p.

Donner, J., Jungner, H. & Kurtén, B. 1979. Radiocarbon dates of mammoth finds in Finland compared with radiocarbon dates of Weichselian and Eemian deposits. Bulletin of the Geological Society of Finland 51, 45–54.

Eriksson, B. 2005. Interglasiaali- ja Interstadiaalivaiheet. In: Johansson, P. & Kujansuu, R. (eds.), Pohjois-Suomen maaperä: maaperäkarttojen 1:400 000 selitys (Summary: Quaternary deposits of Northern Finland – Explanation to the maps of Quaternary deposits 1 : 400 000). Espoo: Geologian tutkimuskeskus, 105–115. [in Finnish]

Forsström, L. 1988. The northern limit of pine forest in Finland during the Weichselian interstadials. Annales Academiae Scientiarum Fennicae, Series A III, Geologica – Geographica 147, 24 p.

Gibbard, P., Forman, S., Salomaa, R., Alhonen, P., Jungner, H., Peglar, S., Suksi, J. & Vuorinen, A. 1989. Late Pleistocene stratigraphy at Harrinkangas, Kauhajoki, western Finland. Annales Academiae Scientiarum Fennicae, Series A III, Geologica – Geographica 150, 36 p.

Heijnis, H. 1992. Uranium/thorium dating of Late Pleistocene peat deposits in N.W. Europe. Groningen: Rijksuniversiteit Groningen, Ph.D Thesis,149 p.

Heijnis, H. & Van der Plicht, J. 1992. Uranium/thorium dating off LLate Pleistocene peat deposits in NW Europe, uranium/thorium isotope systematics and open-system behaviour of peat layers. Chemical Geology (Isotope Geoscience Section) 94, 161–171.

Heikkinen, A. 1975. Radiohiilimenetelmän käytöstä moreenin

sisältämän vanhan hiiliaineksen ajoituksessa (Summary: Radiocarbon dating of old charcoal contained in till). Geologi 27, 51–53. [in Finnish]

Helmens, K., Räsänen, M., Johansson, P., Jungner, H. & Korjonen, K. 2000. The last interglacial-glacial cycle in NE Fennoscandia: a nearly continuous record from Sokli (Finnish Lapland). Quaternary Science Reviews 19, 1605–1623.

Hirvas, H. 1991. Pleistocene stratigraphy of Finnish Lapland. Geological Survey of Finland, Bulletin 354, 123 p.

Hirvas, H., Kujansuu, R. & Tynni, R. 1976. Till stratigraphy in Northern Finland. International Geological Correlation Program, Project 731/1/24: Quaternary Glaciations in the Northern Hemisphere, Report 3. Prague: Bellingham, 256–273.

Hirvas, H., Alfthan, A., Pulkkinen, E., Puranen, R. & Tynni, R. 1977. Raportti malminetsintää palvelevasta maaperätutkimuksesta Pohjois-Suomessa vuosina 1972–1976 (Summary: a report of glacial drift investigations for ore prospecting purposes in northern Finland 1972–1976). Geological Survey Finland, Report of Investigation 19, 53 p. [in Finnish]

Hirvas, H., Lagerbäck, R., Mäkinen, K., Minell, H., Nenonen, K., Olsen, L., Riiber, K., Rodhe, L., Sutinen, R. & Thoresen, M. 1986. Map of Quaternary geology, sheet 4: Quaternary stratigraphy, Northern Fennoscandia, 1:1 milj. Geological Surveys of Finland, Norway and Sweden.

Hütt, G., Punning, J.-M. & Smirnov, A. 1982. The potentiel use of the TL dating method in chronological studies of Late Pleistocene. Pact, Vol. 7, part 1, 1982. Proceedings of the second Nordic Conference on the Application of Scientific Methods in Archaeology, Helsingör, Denmark, 17–19 August 1981, 105–114.

Hütt, G., Jungner, H., Kujansuu, R. & Saarnisto, M. 1993. OSL and TL dating of buried podsols and overlying sands in Ostrobothnia, western Finland. Journal of Quaternary Science 8, 125–132.

Iisalo, E. 1992. Observations on the stratigraphy of Weichselian tills and subtill eskers in central Ostrobothnia, Finland. Geological Survey of Finland, Report of Investigation 112, 42 p.

Ilvonen, E. 1973. Eem-kerrostuma Savukosken Soklilla (Summary: An Eem-Interglacial deposit Sokli in Savukoski, Finnish Lapland). Geologi 25, 81–84. [in Finnish]

Ivanovich, M. & Harmon, R. 1982. Uranium series Disequilibrium, Applications in Geochronology. Oxfort: Clarendon, 571 p.

Jungner, H. 1987. Thermoluminescence dating of sediments from Oulainen and Vimpeli, Ostrobothnia, Finland. Boreas 16, 231–235.

Kankainen, T. & Huhta, P. 1986. Om åldern på organiska avlagringar under morän i Finland (in Swedish). In: 17th Nordiska Geologmötet, Helsingfors Universitet, 12.–15.5.1986 : abstracts. Helsingfors: Helsingfors Universitet, p. 67.

Korpela, K. 1969. Die Weichsel-Eiszeit und ihr Interstadial in Peräpohjola (nördliches Nordfinnland) im Licht von submoränen Sedimenten. Annales Academiae Scientiarum Fennicae, Series A III, Geologica – Geographica 99, 1–108.

Kujansuu, R. 1975. Marrasjärven interstadiaalinen harju Keski-Lapissa (Summary: Interstadial esker at Marrasjärvi, Finnish Lapland). Geologi 27, 45–50. [in Finnish]

Kujansuu, R., Saarnisto, M., Räisänen, M.L. & Hansel, A.K. 1991. Fossil soil of Kärjenkoski and its correlatives in Ostrobothnia, western Finland. In: Autio, S. (ed.), Geological Survey of Finland, Current Research 1989–1990. Geological Survey of Finland, Special Paper 12, 119–126.

Kujansuu, R. & Eriksson, B. 1995. Pre-Late Weichselian subglacial glaciofluvial erosion and accumulation at Vuotso, Finnish Lapland. In: Autio, S. (ed.), Geological Survey of Finland, Current Research 1993–1994. Geological Survey of Finland, Special Paper 20, 75–82.

Lagerbäck, R. 1988. The Veiki moraines in northern Sweden – widespread avidence of an Early Weichselian deglaciation. Boreas 17, 469–486.

Lagerbäck, R. & Robertsson, A.-M. 1988. Kettle holes – stratigraphical archives for Weichselian geology and paleoenvironment in northernmost Sweden. Boreas 17, 439–468.

Lundqvist, J. 1967. Submoräna sediment i Jämtlands län. Sveriges Geologiska Undersökning C 618, 267 p.

Mangerud, J. 1991. The last ice age in Scandinavia. In: Late Quaternary stratigraphy in the Nordic countries 150.000–15.000 B.P. The XXIV Uppsala Symposium in Quaternary Geology. Striae 34, 15–29.

Martinson, D.G., Pisias, N.G., Hays, J.D., Imbrie, J., Moore, T.C. & Shackleton, N.J. 1987. Age dating and the orbital theory of the ice ages; development of a high-resolution 0 to 300,000-year chronostratigraphy. Quaternary Research 27, 1–29.

Mejdahl, V. 1991. A Survey of Archaeological and Geological Samples dated in 1990. Geoskrifter nr. 40. Geologisk Institut, Aarhus Universitet, 35 p.

Mejdahl, V., Shlukov, A.I., Shakhovets, S.A., Voskovskaya, L.P. & Lyashenko, M.G. 1992. The effect of shallow traps: a possible source of error in TL dating of sediments. Ancient TL 10, 22–25.

Mäkinen, K. 1979. Interstadiaalinen turvekerrostuma Tervolan Kauvonkankaalla (Summary: An interstadial peat layer at Kauvonkangas, Finnish Lapland). Geologi 31, 82–87. [in Finnish]

Mäkinen, K. 1981. Glaciofluvial Deposits and their Stratigraphic Position in northern Finland. In: Gorbunov, G.I. (ed.), Glacial Deposits and Glacial History in eastern Fennoskandia, Academy of Sciences of the USSR, Apatity, 63–70.

Mäkinen, K. 1985. On the till-covered glaciofluvial formations in Finnish Lapland. Glaciofluvium. The 16th Uppsala Symposium in Quaternary Geology. Striae 22, 33–40.

Nenonen, K. 1995. Pleistocene stratigraphy and reference sections in southern and western Finland. Geological Survey of Finland, Miscellaneous Publications, Ph.D. Thesis, 205 p.

Niemelä, J. & Jungner, H. 1991. Thermoluminescence dating of late Pleistocene sediments related to till-covered eskers from Ostrobothnia, Finland. In: Autio, S. (ed.), Geological Survey of Finland, Current Research 1989–1990. Geological Survey of Finland, Special Paper 12, 135–138.

Olsen, L., Mejdahl, V. & Selvik, S. 1996. Middle and Late Pleistocene stratigraphy, chronology and glacial history in Finnmark, North Norway. Norges Geologiska Undersökningen, Bulletin 429, 1–111.

Saarnisto, M. & Lunkka, J.P. 2004. Climate variability during the last interglacial-glacial cycle in NW Eurasia. In: Battarbee, R.W., Gasse, F. & Stickley, C.E. (eds.), Past climate variability through Europe and Africa. Developments in Paleoenvironmental Research 6. Dordrecht: Kluwer Academic Publishers, 443–464.

Sarala, P. 2005. Etelä-Lapin glasiaalimorfologinen kehitys Veiksel-jääkauden aikana. In: Ojala, A.E.K. (ed.), Pohjoisten alueiden kvartääritutkimuksen seminaari. Kilpisjärven biologinen asema 13–15.1.2005. Järjestäjä: Suomen Kvartääritutkimuksen Kansalliskomitea. Abstraktikokoelma. Geologian tutkimuskeskus, arkistoraportti, P 22.4.023., 68 p. [in Finnish]

Schulz, H-P., Eriksson, B., Hirvas, H., Huhta, P., Jungner, H., Purhonen, P., Ukkonen, P. & Rankama, T. 2002. Excavations at Susiluola cave (Tiivistelmä: Susiluola). Helsinki: Suomen Museo 2002/109, 5–45.

Siivonen, L. 1975. New results on the history and taxonomy of the mountain, forest and domestic reindeer in Northern Europe. Biological Papers of the University of Alaska, Special Report 1, 33–41.

Sutinen, R. 1982. Glasiaalistratigrafiasta Pudasjärven alueella

Peräpohjolassa (Summary: On glacial stratigraphy in Pudasjärvi area, Peräpohjola). Geologi 34, 191–195. [in Finnish]

Sutinen, R. 1984. On the glacial stratigraphy in Pudasjärvi area, Peräpohja. In: The 15th Uppsala Symposium : ten years of Nordic till research. Striae 20, 91–98.

Sutinen, R. 1992. Glacial deposits, their electrical properties and surveying by image interpretation and ground penetrating radar. Geological Survey of Finland, Bulletin 359, 123 p.

Svendsen J., Alexanderson, H., Astakhov, V., Demidov, I., Dowdeswell, J., Funder, S., Gataullin, V., Henriksen, M., Hjort, C., Houmark-Nielsen, M., Hubberten, H., Ingólfson, O., Jakobsson, M., Kjær, K., Larsen, E., Lokrantz, H., Lunkka, J., Lyså, A., Mangerud, J., Matiouchkov, A., Murray, A., Möller, P., Niessen, F., Nikolskaya, O., Polyak, P., Saarnisto, M., Siegert, C., Siegert, M., Spielhagen, R. & Stein, R. 2004. Late Quaternary ice sheet history of Northern Eurasia. Quaternary Science Reviews 23, 1229–1271.

Ukkonen, P., Lunkka, J.P., Jungner, H. & Donner, J. 1999. New radiocarbon dates from Finnish mammoths indicating large ice-free areas in Fennoscandia during the Middle Weichselian. Journal of Quaternary Science 14, 711–714.

Van der Wijk, A., El-Daoushy, F., Arends, A. & Mook, G. 1986. Dating peat with U/Th disequilibrium; some geochemical considerations. Chemical Geology (Isotope Geoscience Section) 59, 283-292.

Van der Wijk, A. 1987. Radiometric dating by alpha spectrometry on uranium series nuclides. Groningen: Rijksuniversiteit Groningen. Ph.D. Thesis, 155 p.

Van der Wijk, A., Mook. G. & Ivanovich, M. 1988. Correction for environmental 230Th in U/Th disequilibrium dating of peat. The Science of the Total Environment 70, 19–40.

Visocekas, R. 2000. Monitoring anomalous fading of TL of feldspars by using far-red emission as a gauge. Radiation Measurements 32, 499–504.

Vogel, J.C. & Kronfeld, J. 1980. A new method for dating peat. South African Journal of Science 76, 557–558.

Quaternary studies in the northern and Arctic regions of Finland
Edited by Antti E.K. Ojala
Geological Survey of Finland, Special Paper 40, 79– 84, 2005.

DATING OF PALSAS

by
Matti Seppälä

Department of Geography, University of Helsinki, P.O. Box 64, FIN- 00014, FINLAND
E-mail: matti.seppala@helsinki.fi

Seppälä, Matti, 2005. Dating of palsas. *Geological Survey of Finland, Special Paper 40*, 79–84, 13 figures and 1 table.

The dating of palsas is based on changes in ecological conditions caused by the uplift of the mire surface. When a hummock rises, peat formation on the palsa differs from that of the wet mire. Xerophilic peat is formed on the dry palsa surface. Material for dating is sampled from both sides of the contact between peat layers. Surface peat can be eroded by wind-blown ice and snow crystals. If the contact between wet fen peat and xerophilic *Ericales-Bryales* peat is eroded then we cannot date the formation of a palsa. Palsas in Finnish Lapland are mainly less than 1000 years old. Some new palsas have been observed that are only two to four years old on mires with old palsas.

Key words (GeoRef Thesaurus, AGI): palsas, genesis, peat, absolute age, C-14, Holocene, Lapland Province, Finland

Introduction

Radiocarbon dating of organic material is a standard method also often used in palsa studies. Palsas are one of the marginal permafrost features of the permafrost zone (e.g. Zoltai 1971, Seppälä 1988). They are peat hummocks with a permanently frozen core rising above the mire surface. Frozen peat samples can be cored through the palsas and the cores can readily be used to investigate, for example, pollen and peat stratigraphy (e.g. Ruuhijärvi 1962, Salmi 1970, Vorren & Vorren 1975) and to perform macrofossil analyses (e.g. Oksanen et al. 1998, 2001). However, it is then a different matter to determine what these analyses tell us about the date of formation of palsas that have gone through several stages before the present phase of their development.

The main aim of this paper is to discuss, with examples from Finland, the main problems in the sampling of palsas and the interpretation of palsa datings.

Palsa formation

The formation of palsas has been experimentally studied in the field (Seppälä 1982) and observations have also been made of palsas in different stages of formation (Seppälä 1986, 1988).

The rise of a palsa hummock leads to changes in the ecological conditions of the plants forming the peat (Figs. 1 and 2). The wet palsa surface dries and new plants colonize the rising hummock. In this stage, a xerophilous peat layer covers the uppermost flark peat (e.g. Vorren & Vorren 1975).

Samples for dating the initial stage of palsa formation should be collected from the contact between these two peat layers (Vorren 1972, Seppälä 1988).

Fig. 1. Some new 1.5-year-old palsas on Vaisjeäggi palsa mire, Utsjoki. (Photo: M. Seppälä, 2003)

Fig. 3. A mature palsa in Iitto, Enontekiö. (Photo: M. Seppälä, 1996)

Fig. 2. A new some-years-old palsa on Luovdijeäggi palsa mire, western Utsjoki. (Photo: M. Seppälä, 1999)

Fig. 4. An uncovered palsa surface on Vaisjeäggi palsa mire, Utsjoki. (Photo: M. Seppälä, 1994)

Fig. 5. Remains of a former palsa with a peat rim, Abisko, northern Sweden. (Photo: M. Seppälä, 1985)

Sphagnum, *Carex* and *Eriophorum* species grow on the wet mire surface. The conditions are similar to string (aapa) mires.

Shrubs grow on the uplifted, mature palsa surface (Fig. 3), such as *Ericales* species, *Bryales* mosses and *Cladina* lichens, which form poorly humified peat at a very slow rate.

Problems arise in dating when the palsa surface is subject to abrasion. In the winter, when palsa surfaces are uncovered from the snow, ice crystals blown by the wind (e.g. Seppälä 2004) abrade the peat (Fig. 4), and fen peat forms the present palsa surface. Dating of the surface peat does not tell us the age of the palsa.

The further development of a palsa can occur through block erosion, when large pieces of peat slide down the steep edge of the frozen core and the palsa loses its insolation and starts to melt and collapse. As a result a peat rim is formed, which surrounds a wet pool or new growth of flark (Fig. 5).

Palsas sometimes lose their frozen core as a result of melting from the bottom, and the palsa slowly sinks into the wet mire surface (Fig. 6).

A palsa is an ecological island that is under continuous development. This should be kept in mind as the basis for the dating of palsas.

Fig. 6 . A sinking palsa that is thawing from the bottom. Vaisjeäggi palsa mire, Utsjoki. (Photo: M. Seppälä, 1985)

Fig. 7. Abraded mature palsas on Luovdijeäggi palsa mire, western Utsjoki. In the foreground some pounus and a thermokarst pond. (Photo: M. Seppälä, 1999)

Examples of palsa datings in Finland

Salmi (1972) reported two dates, 4150 ± 50 B.P. and 3650 ± 170 B.P, for the surface peat of palsas from western Finnish Lapland. However, these datings do not tell the age of the palsa formation. We have dated several peat samples from one mire, Vaisjeäggi, in northern Lapland and the dates ranged from 840 to 1630 B.P. (Table 1). The age of peat depends on the amount of abrasion. On Luovdijeäggi mire the present abrasion of peat is very intensive, c. 40 cm peat eroded (Seppälä 2003). Most of the palsa surfaces are not covered by vegetation (Fig. 7). Surface peat of one palsa was dated to 2710±70 B.P. (Fig. 8) and the contact of the xerofilic peat and the fen peat could be found from an erosion remnant and the dated peat above the contact was only 380±60 B.P. while the fen peat below the contact gave 930±70 B.P. (Seppälä 2003, Fig. 7). There probably has happened some early abrasion before the final deposition of xerofilic peat on the palsa. Normally the age differences between these contact samples are not so big (Fig. 9).

If we take only one sample from the contact layer we cannot be so sure when the palsa was in fact formed (Fig. 10). The sample will be a mixture of fen peat and xerofilic surface peat formed on the palsa.

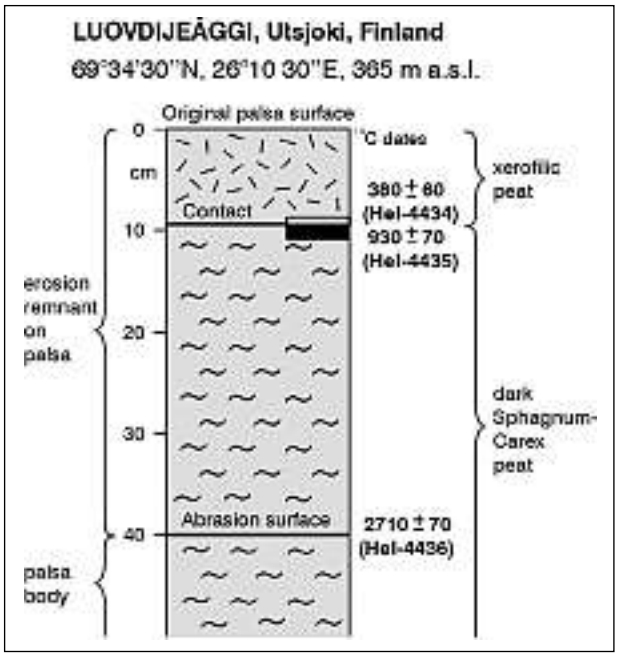


Fig. 8. Dating of the contact layer of fen peat and xerofilic peat in a erosional remnant on an abraded palsa and dating of peat of the abrasion surface. (Seppälä 2003).

Table 1. Dated surface peat samples from palsas on Vaisjeäggi, Utsjoki, Finland. 0.5 cm thick layers of peat sampled from abrasion surfaces of four different palsas.

14C ages BP	Lab. code	calAD (Intercal98) (1sigma)
840 ± 100	(Hel-1041)	1160–1280
1070 ± 90	(Hel-1040)	860–1040
1210 ± 110	(Hel-1038)	760–900
1630 ± 130	(Hel-1039)	320–560

Peat on a silt-core palsa can be reasonably old, as in Munnikurkkio, Enontekiö, where we have probably the highest palsas in Finland (7 to 8 m in height) (Fig. 11). There, a peat sample c. 30 cm below the surface gave an age of 7740 ± 260 B.P. The peat lying 50–70 cm below the surface is often about 6000 radiocarbon years old (Figs. 11 and 12).

The ages of the bottom layers of organic deposits under the palsas often range from 8000 to 9700 B.P. (Fig. 11, Seppälä 1971), but they tell nothing of the

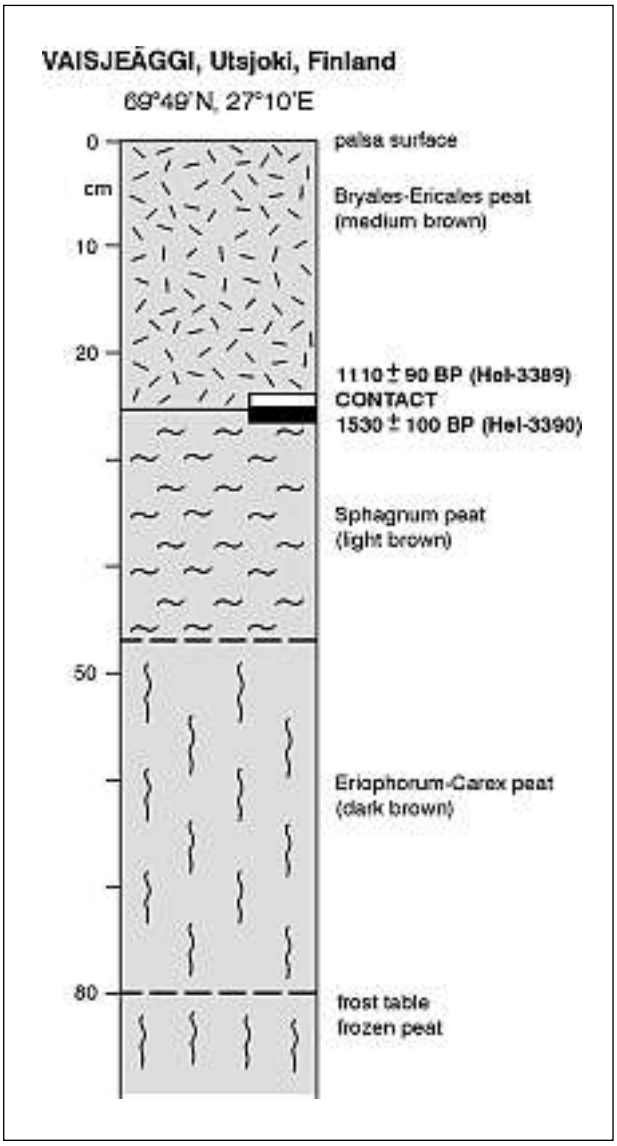


Fig. 9. An example of palsa dating from the contact layer of different peats.

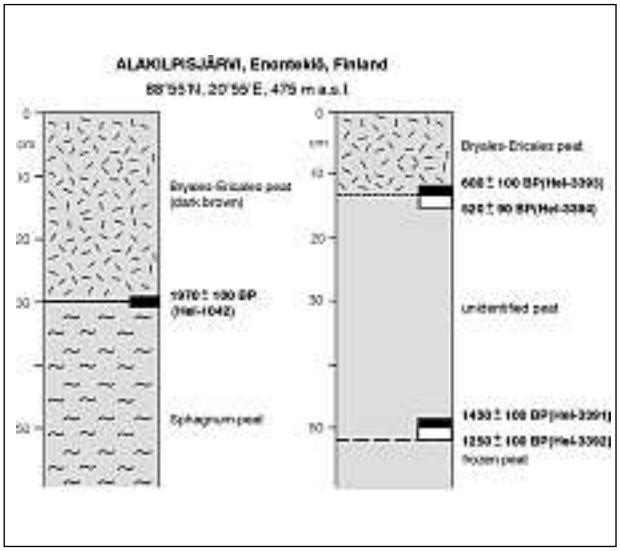


Fig. 10. Datings of two palsa exposures with contact layers.

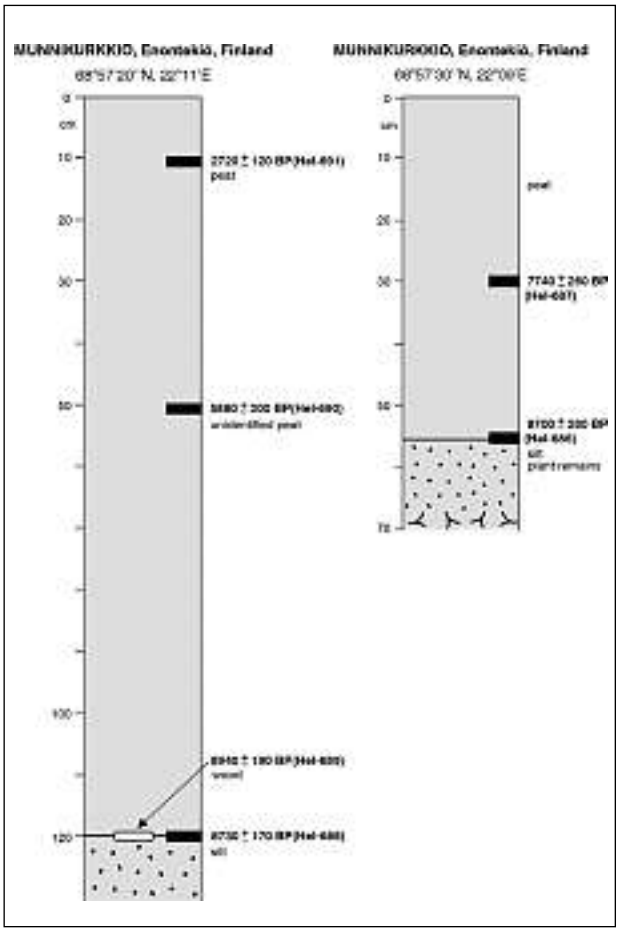


Fig. 11. Dated palsa material from two silt-cored palsas.

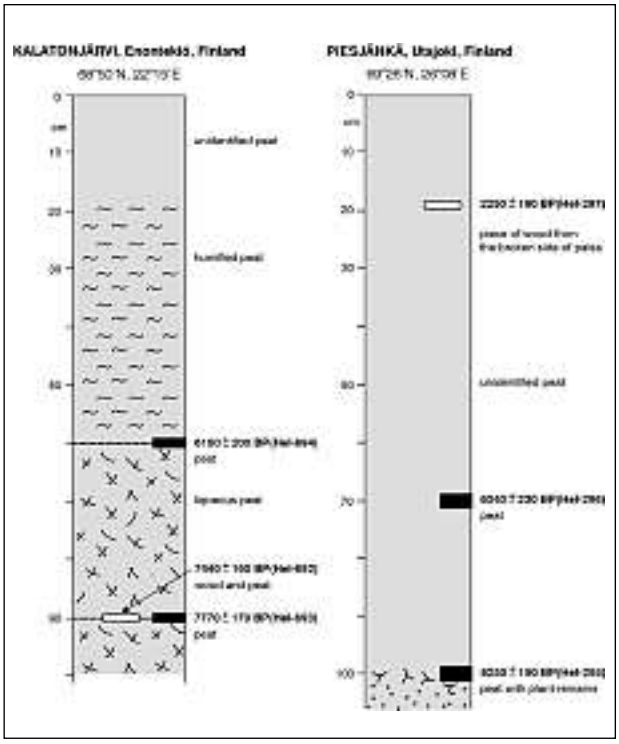


Fig. 12. Two dated palsa sections with rather old peat. No dates from the contact layers.

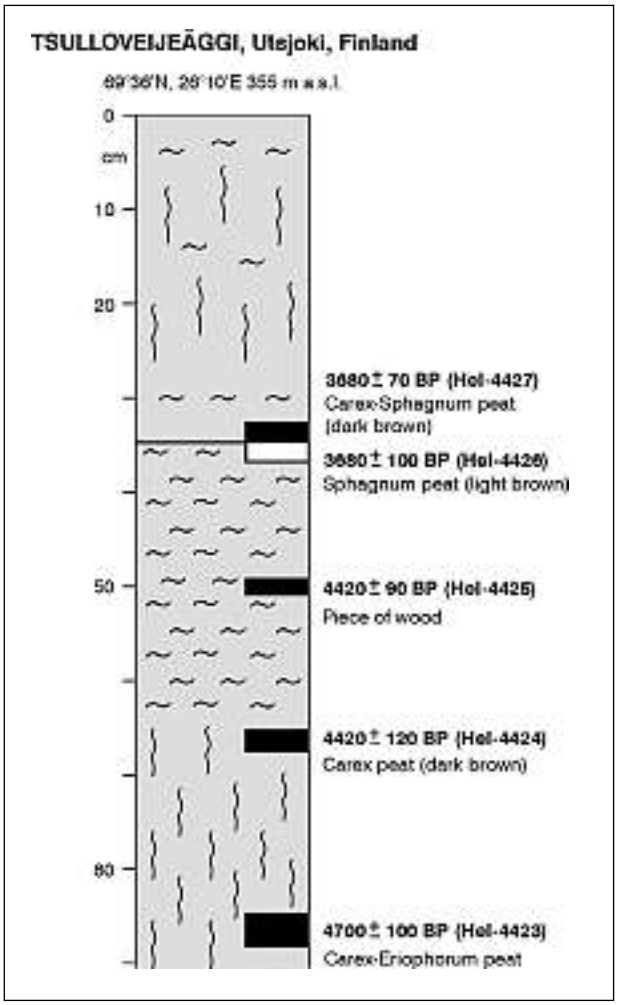


Fig. 13. Stratigraphy of dated peat from the edge of a very large palsa. The sampled layers have probably been tilted by the frost heave and therefore the samples have similar ages.

age of palsas and merely correspond to the beginning of the peat formation.

During the thawing of a palsa the wet peat is completely mixed and its strata are destroyed. If a new palsa is formed in this type of location, the material is not suitable for dating or pollen analyses.

We sampled one good exposure on the edge of a very large palsa about 5 m in height in Tsulloveijeäggi, western Utsjoki. The five datings differed unusually little (Fig. 13). The uppermost date from the depth of 35 cm was 3680 ± 70 B.P. and the two samples from the depths of 50 and 65 cm gave almost exactly the same date of 4420 B.P., while from 85 cm depth a date of 4700 ± 100 B.P was determined. An explanation for the dates being so close might be that the samples were taken from a steep edge and the peat layers, even though vertically distant from each other, had inclined strongly when the palsa was formed and rose from the mire surface.

Discussion and conclusions

This article illustrates that we have to be very careful in sampling when dating palsas. We should consider the processes affecting palsa formation and finally make interpretations according to this knowledge. If we can identify the contact between fen peat and xerofilic peat above it then we have a possibility to date the rising of the peat hummock from the wet mire surface.

By using pollen analyses, Ruuhijärvi (1962) concluded that palsa formation started in the Subatlantic period after 5000 years B.P. This seems to be much too early. Vorren (1972, 1979) carried out careful stratigraphic studies of palsas in northern Norway. At first, Vorren (1972) proposed that the transition to ombrogenous peat deposition probably took place between 3000 and 2500 B.P. Three more dates from Norwegian palsas not far from Finland gave ages of about 1000 BP to 450 B.P. (Vorren 1979). According to our observations, palsas in Finnish Lapland are mainly less than 1000 years old. Palsas are currently forming in Finnish Lapland and we have small palsas that are only some years old; if the conditions remain suitable they will grow larger. Old palsas often disappear because of strong winter storms that abrade the palsa surfaces (Seppälä 2003, 2004) and as a result of the cyclic development of palsas (Seppälä 1986, 1988).

REFERENCES

Oksanen, P.O., Kuhry, P., Alekseeva, R.N. & Kanev, V.V. 1998. Permafrost dynamics at the Rogovaya River peat plateau, Subarctic Russia. In: Lewkowicz, A.G. & Allard, M. (eds.), Permafrost, Seventh International Conference, Yellowknife, Canada, Proceedings, Collection Nordicana 57, 847–854.

Oksanen, P., Kuhry, P. & Alekseeva, R.N. 2001. Holocene development of the Rogovaya River peat plateau, European Russian Arctic. The Holocene 11, 25–40.

Ruuhijärvi, R. 1962. Palsoista ja niiden morfologiasta siitepölyanalyysien valossa. (Zusammenfassung: Über die Palsamoore und deren Morphologie im Lichte der Pollenanalyse). Terra 74, 58–68.

Salmi, M. 1970. Investigations on palsas in Finnish Lapland. In: Ecology of the Subarctic regions, Proceedings of the Helsinki Symposium 1966, UNESCO, Paris, 143–153.

Salmi, M. 1972. Present development stages of palsas in Finland. In: Virgin peatlands. The Proceedings of the 4th International Peat Congress, Otaniemi, Finland, Volume 1, 121–141.

Seppälä, M. 1971. Evolution of eolian relief of the Kaamasjoki-Kiellajoki river basin in Finnish Lapland. Fennia 104, 1–88.

Seppälä, M. 1982. An experimental study of the formation of palsas. The Roger J.E. Brown Memorial Volume. Proceedings Fourth Canadian Permafrost Conference, Calgary, Alberta. National Research Council of Canada, 36–42.

Seppälä, M. 1986. The origin of palsas. Geografiska Annaler 68A, 141–147.

Seppälä, M. 1988. Palsas and related forms. In: Clark, M.J. (ed.),

Advances in Periglacial Geomorphology. Chichester: John Wiley & Sons Ltd, 247–278.

Seppälä, M. 2003. Surface abrasion of palsas by wind action in Finnish Lapland. Geomorphology 52, 141–148.

Seppälä, M. 2004. Wind as a geomorphic agent in cold climates. Cambridge: Cambridge University Press, 358 p.

Vorren, K.-D. 1972. Stratigraphical investigations of a palsa bog in northern Norway. Astarte 5, 39–71.

Vorren, K.-D. 1979. Recent palsa datings, a brief survey. Norsk geografisk Tidsskrift 33, 217–220.

Vorren, K.-D. & Vorren, B. 1975. The problem of dating a palsa. Two attempts involving pollen diagrams, determination of moss subfossils, and C14-datings. Astarte 8, 73–81.

Zoltai, S.C. 1971. Southern limit of permafrost in peat landforms, Manitoba and Saskatchewan. The Geological Association of Canada, Special Paper 9, 305–310.

Quaternary studies in the northern and Arctic regions of Finland
Edited by Antti E.K. Ojala
Geological Survey of Finland, Special Paper 40, 85– 90, 2005.

QUANTITATIVE ANALYSIS OF PAST AND PRESENT POLLEN-CLIMATE AND POLLEN-VEGETATION RELATIONSHIPS IN THE NORTHERN BOREAL FOREST ZONE OF FINNISH LAPLAND, A PROGRESS REPORT

by
Mari Hagberg

Department of Geosciences, Linnanmaa (P.O. Box 3000), FIN-90014, University of Oulu, FINLAND
E-mail:mari.hagberg@oulu.fi

Hagberg, Mari, 2005. Quantitative analysis of past and present pollen-climate and pollen-vegetation relationships in the Northern Boreal forest zone of Finnish Lapland, a progress report. *Geological Survey of Finland, Special Paper 40*, 85–90, 5 figures and 1 table.

The pollen-climate and pollen-vegetation relationships at Nellim, in northern Finnish Lapland, are examined at a high spatial and temporal resolution and in a quantified manner. The peat profile to be analyzed is dated by AMS 14C-analyses of solely Sphagnum remains and wiggle-matching the results with the atmospheric 14C curve. Sampling for pollen analysis is at a near-annual resolution on the basis of the age-depth chronology. The study also incorporates the latest developments in modelling pollen loading and reconstructing the vegetation coverage around the site being studied. Preliminary simulation results suggest that the relevant source area of pollen in the Nellim pine forests is about 1 km. This indicates that a reconstruction of the vegetation is possible within a 1 km radius around the study site. Further research is still required since the simulations are made using pollen productivity estimates from Denmark and southern Sweden which will, in all likelihood, be different from those in northern Finland.

Key words (GeoRef Thesaurus, AGI): paleoecology, paleobotany, peat, pollen analysis, pollen influxes, models, Holocene, Inari, Finland

Introduction

The purpose of this study is to analyse pollen-climate and pollen-vegetation relationships quantitatively at a high spatial and temporal resolution, in order to reconstruct climate and human-induced changes in the vegetation of the Nellim area during the last 300 – 500 years.

This study concentrates especially on Scots pine (*Pinus sylvestris* L.) (Hämet-Ahti et al. 1998) since the study area in northern Finnish Lapland is situated in the middle of pine dominated forest and is north of the forest limit of Norway spruce (*Picea abies* (L.) H. Karst.) (Hämet-Ahti et al. 1998). Pine pollen is also suitable for investigating whether a temperature signal can be attained from peat since earlier studies have shown that the pollen production of Scots pine correlates with July temperature of the last growing season (Autio & Hicks 2004). The research is closely related to an EU 5FP project called PINE (Predicting Impacts on Natural Ecotones, Contract no: EVK2-CT-2002-00136) which focuses on northern boreal vegetation and produces similar data from other parts of the pine forest belt.

A peat profile for pollen analysis was taken close to the centre of the village of Nellim in the administrative district of Inari. The site was chosen to be close to the village centre so that the effect of human influence

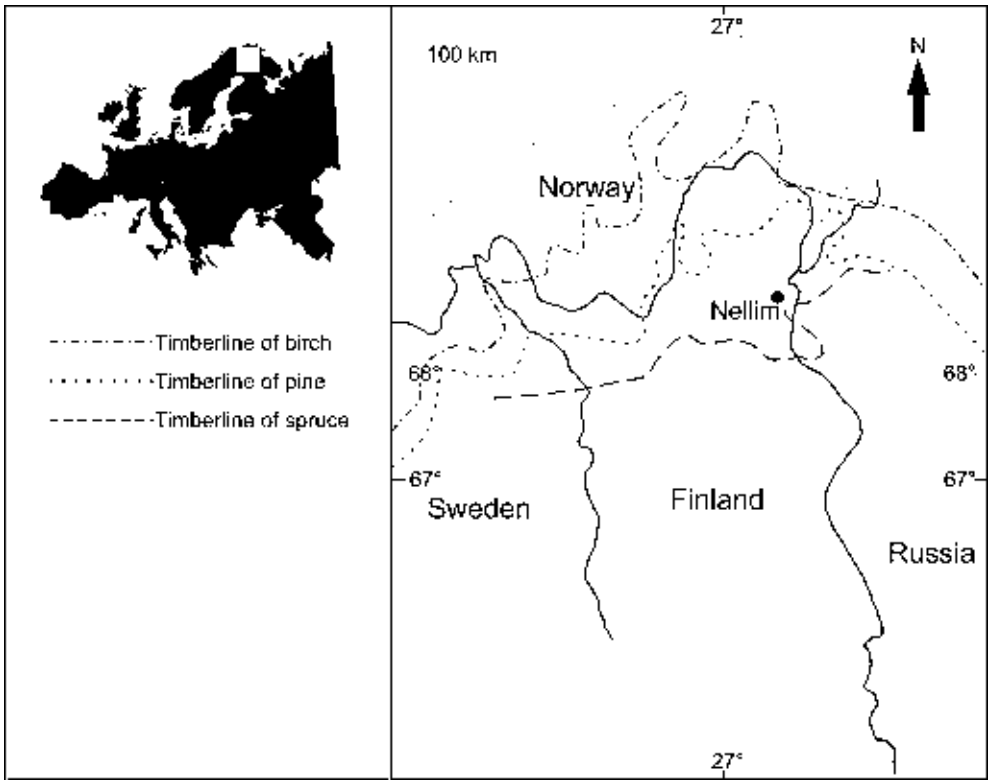


Fig. 1. The study area and location of the latitudinal forest limits of the main tree species in Finnish Lapland.

on the vegetation could be detected in the pollen assemblages. Nellim is also a focus for socio-economic research within the PINE project and it, additionally has a continuous record of annual pollen deposition monitored by modified Tauber traps (Tauber 1974, Hicks & Hyvärinen 1986) since 1983 (Fig. 1).

Throughout, the research aims at quantification, a high spatial resolution and a near-annual temporal resolution. To this end, in the pollen analysis pollen accumulation rates (PAR; number of palynomorphs per square cm per year) are used instead of the traditional percentage representation. PAR's enable a better explanation of the vegetation on site since the representation of each species is unaffected by the changes in the abundance of other species (Davis & Deevey 1964). Especially in areas where a high pollen producing species, such as Scots pine, is so abundant, using PAR's insures that the changes happening in the abundance of low pollen producing species can also be detected. With the percentage manner of presentation pine simply swamps everything.

Along with the pollen analysis I am using different models which describe the spread and accumulation of pollen (Parsons & Prentice 1981, Prentice & Parsons 1983, Sugita 1994). These models can be used to reconstruct the quantitative vegetation coverage around the sample site from the fossil pollen assemblages (Sugita et al. 1999, Nielsen & Odgaard 2005).

Materials and methods

For the pollen analysis a peat profile was collected with a spade near the centre of the village of Nellim. The sample site was chosen so that it is possible to detect the anthropogenic influence on the vegetation resulting from the development of the village through changes in the pollen assemblages. Previous work has shown that, in this type of environment, changes are only visible within 100 to 200 m of where they occur (Hicks 1993).

In order to construct a robust age-depth model for the peat profile AMS ^{14}C-analyses were conducted on pure *Sphagnum* remains. *Sphagnum* is a moss and because it grows from the same apex all the time the amount of redeposited older material is minimised by using only its' leaves and stems for dating. In order to obtain high-resolution results the samples for dating were taken from slices less than 5 mm thick and pure *Sphagnum* was picked by hand and cleaned from all other plant remains such as rootlets etc. which could affect the results. There were seven samples in total and they were taken at depths of 2.5, 3.5, 5, 7, 9, 11, and 13 cm. The AMS ^{14}C analyses were carried out at the Poznań Radiocarbon Laboratory and the results compared to the atmospheric ^{14}C concentrations in pre- and post-bomb periods. An age-depth chronology (Fig. 2) was then constructed with PozCal software

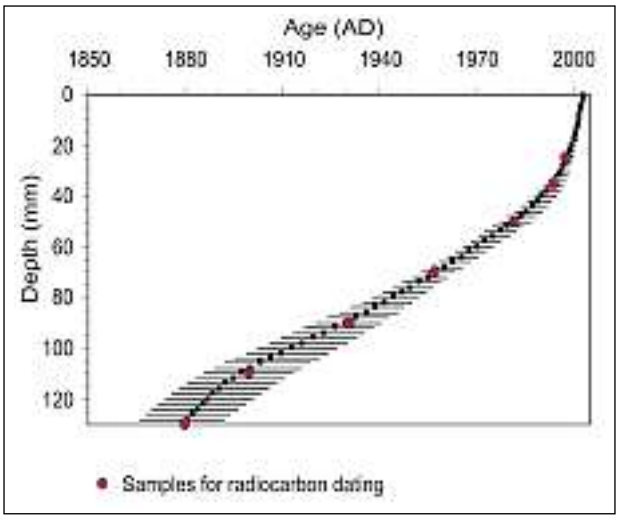


Fig. 2. The age-depth chronology of the Nellim peat profile.

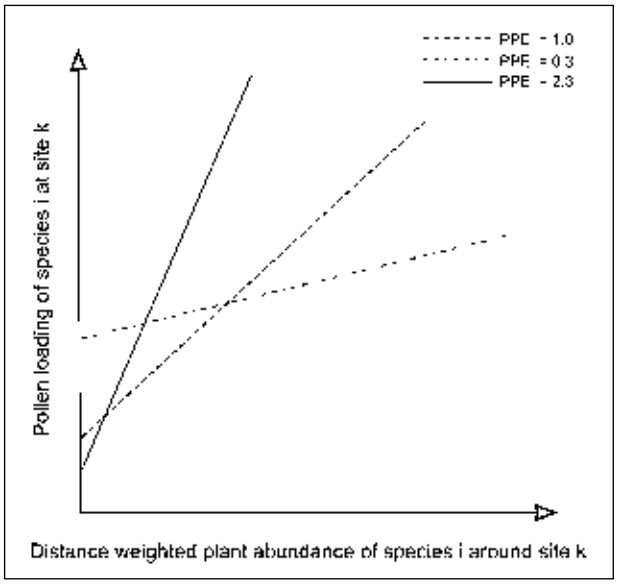


Fig. 3. The linear relation of pollen loading to the distance weighted plant abundance according to equation 1. The offset of the graph from the origin depicts the proportion of background pollen.

which uses a standard algorithm for calibration of ^{14}C ages (Goslar et al. in press).

Using the age-depth chronology, pollen samples were cut from a frozen 4x4 cm block such that the thickness of each sample was as close to one year as possible. Pollen preparation followed a standard procedure of heating for 10 minutes in 10 % KOH followed by acetolysis for 2 minutes with staining and mounting in silicone oil. A known number of *Lycopodium* tablets (Stockmarr 1971, 1972) were added to the samples before preparation so that it was possible to calculate the pollen concentrations and accumulation rates. A minimum number of 500 arboreal pollen grains were counted for each sample. The counts were made in such a way that always an entire slide or a half slide was counted so that the sorting of the grains would not affect the results (see explanation in Hicks 2001). Results are presented as PAR's (grains cm^{-2} $year^{-1}$). The confidence limits on such calculations depend on the relationship between the number of Lycopodium spores added and the number counted (Maher 1980, 1981) and also on the accuracy of the age-depth chronology. Attention has been paid to both of these factors.

The validity of the annual PARs calculated from the peat will be tested by comparing the results with the monitored annual pollen deposition from a pollen trap within the same vegetation belt (N7 in Hicks 2001). A complete 1:1 agreement is not expected since, although the peat slices were cut to represent a single year, this may not be a single flowering season but could be part of one season and part of the following one.

The Extended R-Value (ERV) model relates pollen percentages to vegetation composition within a certain area, by taking the pollen productivities and fall speed of different taxa into account. (Parsons & Prentice 1981, Prentice & Parsons 1983, Sugita 1994) The ERV model incorporates a background component for each species to account for pollen originating from outside the area of the vegetation survey.

The basic assumption of the ERV model is that the pollen loading of taxon i at site k ($y_{i,k}$) is linearly related to the distance-weighted plant abundance ($\psi_{i,k}$) of taxon i around site k:

$$y_{i,k} = \alpha_i \cdot \psi_{i,k} + \omega_i \qquad (1)$$

where α_i is the pollen productivity of taxon i and ω_i is the background pollen loading of taxon i, both of which are taxon-specific constants (Fig. 3) (Prentice & Parsons 1983).

With the ERV models and the HUMPOL software package developed in POLLANDCAL network (NorFA), it is possible to make quantitative reconstructions of vegetation around the study site. Within the network there are several ongoing studies utilising the method and so far the results have been convincing, see for example Sugita et al. (1999), Nielsen & Odgaard (2005). The models enable the delineation of a relative source area of pollen (RSAP) which marks the radius of the local pollen loading and gives the spatial resolution for the vegetation reconstruction.

To calculate the pollen loading within a vegetation community that resembles that of Nellim a 50 km by 50 km simulated landscape with comparable vegetation

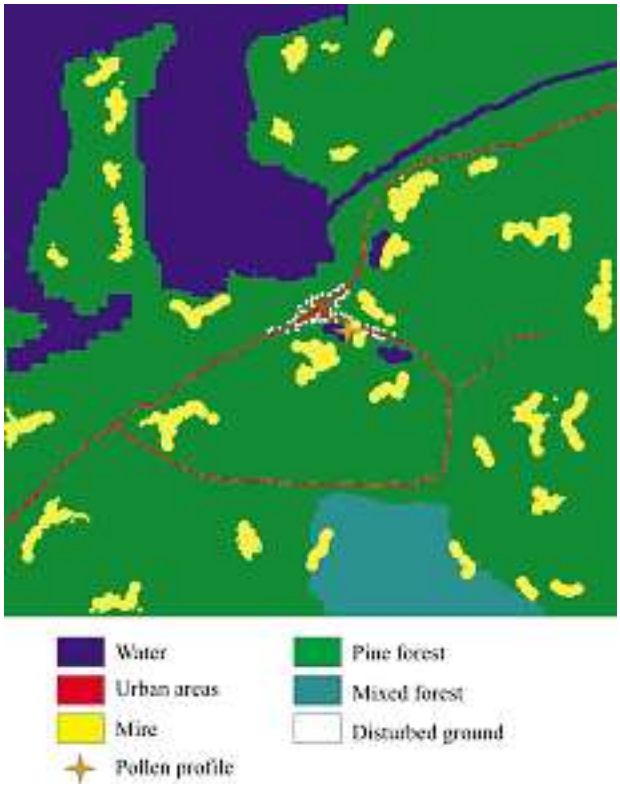


Fig. 4. 50 km x 50 km simulated landscape around Nellim created with HUMPOL Package software Mosaic v 3.0. (Middleton & Bunting 2004). Orange star marks the site of the pollen profile.

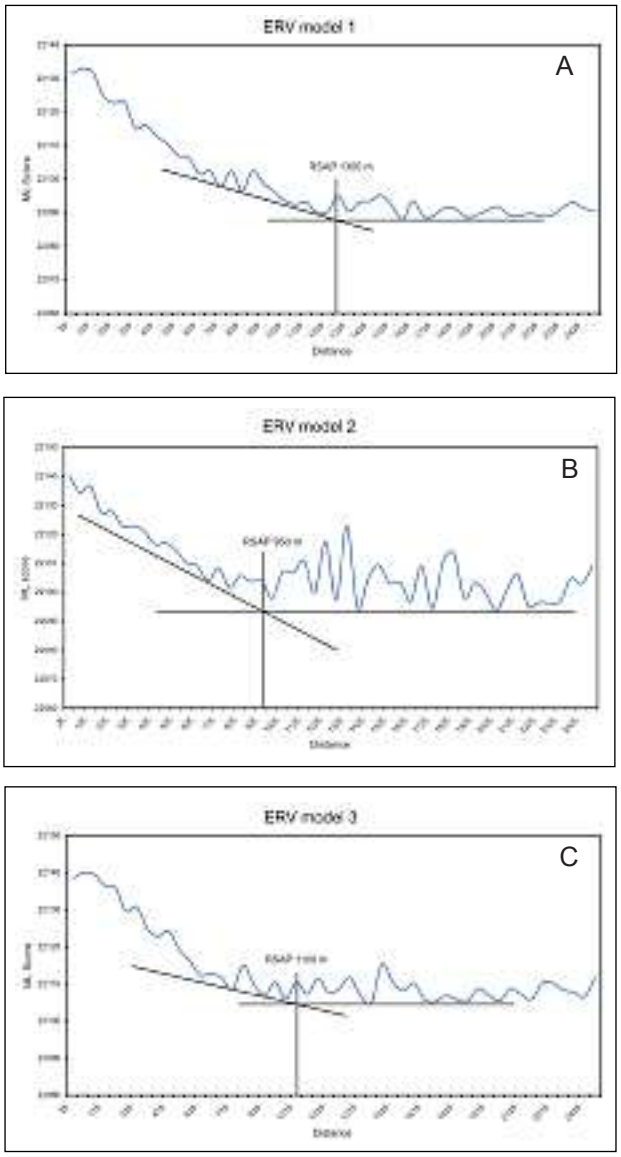


Fig. 5 a-c. Maximum Likelihood Scores (MLS) relative to distance from sample point derived from ERV models 1, 2, and 3 and the Relevant Source Areas of Pollen (RSAP's).

communities was created using the HUMPOL Package software Mosaic v 3.0 (Middleton & Bunting 2004) (Fig. 4, Table 1). For calculating the pollen loading at selected sampling points (23 in total) the simulated landscape was loaded into PolFlow software along with files containing information about the composition of the vegetation units and Pollen Productivity Estimates (PPE). The pollen productivity estimates

Table 1. Vegetation communities and their composition in the simulated landscape.

	Mire (yellow)	Pine forest (green)	Mixed forest (bluish green)	Disturbed ground (white)	Water (blue)	Urban areas (red)
Pinus	0%	75%	40%	0%	non-pollen producing	non-pollen producing
Picea	0%	0%	50%	0%		
Betula	0%	10%	10%	0%		
Juniperus	0%	5%	0%	0%		
Salix	5%	0%	0%	35%		
Calluna	5%	10%	0%	0%		
Poaceae	15%	0%	0%	45%		
Cyperaceae	75%	0%	0%	0%		
Filipendula	0%	0%	0%	20%		

used in the simulation were from southern Sweden and Denmark (Broström et al. 2004, Nielsen & Odgaard 2005). For the calculation of Relevant Source Area of Pollen (RSAP) the HUMPOL Package has PolERV software, which uses the pollen loadings and vegetation abundances obtained from PolFlow to run the ERV models 1, 2 and 3. From the PolERV output calculated maximum likelihood scores can be plotted against the distance from the sample point in order to delimit the extent of the Relevant Source Area of Pollen (Fig. 5).

Results

The pollen counts from the Nellim peat profile are still at a relatively early stage. So far 29 samples have been analysed and their time span goes back to the mid 1950's. Since the amount of data will increase, it is not practical to publish a partial diagram here. Neither can the pollen-climate calibration be attempted at this stage. A complete diagram will be published later in another paper.

Some idea of the spatial extent of the pollen profile can be made, however. Using the preliminary simulation based on an air photo image of the landscape around Nellim the Relevant Source Area of Pollen (RSAP) obtained from the ERV model's maximum likelihood function score (MLS) when it reaches asymptote plotted against the distance from sample point is 1300 m for model 1. ERV model 2 gives some fluctuation in the MLS and reaches the asymptote at 950 m distance while ERV model 3 gives a RSAP of 1100 m (Fig. 5 a–c).

The simulation results suggest that the spatial resolution of the pollen data is about 1 km in the Nellim area. In other words the spatial reconstruction of the vegetation seems to be possible to 1 km distance from the sample point. However, it would be rash to make any firm conclusions from these results since the PPE's used in the modelling are from southern Sweden and Denmark and they are likely to be a different from the actual PPE's of northern Finland. To date PPE's have not been calculated for northern Finland but these should be available in the near future.

Discussions and plans for the future

Models are an effective way to test different scenarios. They are especially useful when the aim is to simulate radical changes and extremities since these are impossible to implement in 'real life'. As shown above the modelling can also give some insight into the limitations and possibilities of the research data. In view of this and comparing the results from the Nellim simulations with Sugita's (1994) simulation results the radius of RSAP in Nellim is a little bit higher than those calculated by Sugita for theoretical landscapes in the Midwest of the United States.

The question remains, however, of what is the effect on the model of having such a very restricted flora, with Scots pine being dominant both inside and outside the RSAP. Since Scots pine is a high pollen producer this may cause some problems and it is clear that this aspect must be studied further. This also highlights the importance of obtaining PPE's which are accurate for northern latitudes.

The investigations will continue with defining the PPE's for northern Finland. It is also important to know the overall amount of background pollen in the study area so that the fossil pollen record can be used effectively to reconstruct the vegetation around the site. Plans have been made to sample large lakes in the study area in order to analyse the proportion of current regional i.e. the background pollen from the sediment surface samples. Lakes have to be chosen in a way that they have a radius larger than the theoretical RSAP so that the pollen loading will represent the amount of pollen coming from outside the RSAP (Sugita 1994). With these results the overall picture of the relationship between pollen and vegetation and pollen and climate in Nellim will most certainly become clearer.

Acknowledgements

First I would like to express my thanks to my supervisors Sheila Hicks and Antti Huusko for introducing me to this fascinating line of study and for useful comments and discussions concerning this work. I also address my gratitude to Matti Eronen and Matti Saarnisto for their comments on earlier version of the manuscript. This paper is a contribution to the POLLANDCAL (POLlen-LANDscape CALibration) network sponsored by NorFA (Nordic Council of Advanced Studies) and co-ordinated by M.-J. Gaillard (University of Kalmar, Sweden) (network website: www.geog.ucl.ac.uk/ecrc/pollandcal/). I am very thankful to all POLLANDCAL members for useful and inspiring discussions during the numerous network workshops held between 2002 and 2004 (2005), and in particular to Shinya Sugita (University of Minnesota, USA) for providing us with the basic research approaches of POLLANDCAL, and for allowing access to his numerous computer programs, and to Jane Bunting and Dick Middleton (University of Hull, U.K.) for developing excellent user-friendly software (the HUMPOL suite of programs) for use in network research activities. The basis for the map in Figure 1 was prepared by

Dmitri Mauquoy. This is also a contribution to the projects PINE and NORDPINE. In this connection grateful acknowledgement is made to Academy of Finland for grant number 1205995.

REFERENCES

Autio, J. & Hicks, S. 2004. Annual variations in meteorological conditions and pollen deposition on the fell Aakenustunturi in northern Finland. Potential for using pollen as a climate proxy. Grana 43, 31–47.

Broström, A., Sugita, S. & Gaillard, M.-J. 2004. Pollen productivity estimates for reconstruction of past vegetation cover in the cultural landscape of Southern Sweden. *The Holocene* **14,** 371–384.

Davis, M.B. & Deevey, E.S. Jr. 1964. Pollen accumulation rates: Estimates from late-glacial sediment of Roger Lake. Science 145, 1293–1295.

Goslar, T., Van der Knaap, W.O., Hicks, S., Andric, M., Czernik, J., Goslar, E., Räsänen, S. & Hyötylä, H. 14C dating of modern peat profiles: pre- and post-bomb 14C variations in the construction of age-depth models. Radiocarbon, in press.

Hicks, S. 1993. Pollen evidence of localized impact on the vegetation of northernmost Finland by hunter-gatherers. Vegetation History and Archaeobotany 2, 137–144.

Hicks, S. 2001. The use of annual arboreal pollen deposition values for delimiting tree-lines in the landscape and exploring models of pollen dispersal. Review of Palaeobotany and Palynology 117, 1–29.

Hicks, S. & Hyvärinen, V.-P. 1986. Sampling modern pollen deposition by means of 'Tauber traps': some considerations. Pollen et Spores 28, 219–242.

Hämet-Ahti, L., Suominen, J., Ulvinen, T. & Uotila, P. (eds.) 1998. Retkeilykasvio (Field Flora of Finland). Helsinki: Finnish Museum of Natural History, Botanical Museum, 656 p.

Maher, L.J. Jr. 1980. The confidence limit is a necessary statistic for relative and absolute pollen data. Proceedings IV International Palynological Conference, Lucknow (1976–77), 152–162.

Maher, L.J. Jr. 1981. Statistics for microfossil concentration measurements employing samples spiked with marker grains. Review of Palaeobotany and Palynology 32, 153–191.

Middleton, R. & Bunting, M.J. 2004. Mosaic v1.1: landscape scenario creation software for simulation of pollen dispersal and deposition. *Review of Palaeobotany and Palynology* **132,** 61–66.

Nielsen, A.B. & Odgaard, B.V. 2005. Reconstructing land-cover from pollen assemblages from small lakes in Denmark. *Review of Palaeobotany and Palynology* **133,** 1–21.

Parsons, R.W. & Prentice, I.C. 1981. Statistical approaches to R-values and pollen-vegetation relationship. Review of Palaeobotany and Palynology **32,** 127–152.

Prentice, I.C. & Parsons, R.W. 1983. Maximum likelihood linear calibration of pollen spectra in terms of forest composition. Biometrics **39,** 1051–1057.

Stockmarr, J. 1971. Tablets with spores used in absolute pollen analysis. Pollen et Spores 13, 615–621.

Stockmarr, J. 1972. Determination of spore concentration with an electronic particle counter. Danmarks Geologiske Undersøgelse Årbok, 87–89.

Sugita, S. 1994. Pollen representation of vegetation in Quaternary sediments: theory and method in patchy vegetation. Journal of Ecology **82**, 881–897.

Sugita, S., Gaillard, M.-J. & Broström, A. 1999. Landscape openness and pollen records: a simulation approach. The Holocene 9, 409–421.

Tauber, H. 1974. A static non-overload pollen collector. New Phytologist 73, 359–369.

Quaternary studies in the northern and Arctic regions of Finland
Edited by Antti E.K. Ojala
Geological Survey of Finland, Special Paper 40, 91– 95, 2005.

A PROPOSAL FOR FORMAL LITHOSTRATIGRAPHICAL NAMES IN THE SUUPOHJA REGION, WESTERN FINLAND

by
Reijo Pitkäranta

Department of Geology, 20014, University of Turku, FINLAND
E-mail: reijo.pitkaranta@utu.fi

Pitkäranta, Reijo, 2005. A proposal for formal lithostratigraphical names in the Suupohja region, western Finland. *Geological Survey of Finland, Special Paper 40*, 91–95, 2 figures and 1 table.

The Suupohja region in southern Ostrobothnia, western Finland, has diverse Quaternary lithostratigraphy. Former and recent observations of the lithostratigraphical units have been compiled and formal names for them proposed. Altogether seventeen identified units (excluding the Holocene clay, peat and eolian deposits) are included to the Suupohja Group. It consists of the following formations, from bottom to top: Karhukangas Formation (nine different lithofacies observed), Kankalo Sand (two or three different lithofacies), Kariluoma Till, Harrinkangas Formation, Kauhajoki Till and Lumikangas Formation. In addition, to the Paleozoic lithified sandstone, which locally has been observed at the bottom above the crystalline Precambrian bedrock, the name "Lauhanvuori Sandstone" is proposed. Majority of the Quaternary sediments have been deposited in glacial (till) or ice-marginal environment (glaciofluvial and glaciolacustrine/-marine sediments). In the upper parts also shore and eolian deposits can be observed. The sediments below the surficial Kauhajoki Till are older than the Middle Weichselian Substage, and the oldest sediments in the deeper parts may originate from the Middle or even Early Pleistocene Epoch.

Key words (GeoRef Thesaurus, AGI): glacial geology, sediments, lithostratigraphy, stratigraphic units, lithofacies, Quaternary, Pleistocene, South Ostrobothnia, Finland

Introduction

The southern part of Ostrobothnia, the Suupohja region, has diverse Quaternary lithostratigraphy with several clearly distinguishable sedimentary units. These lithostratigraphical features have been described in several papers since the 1970's (see e.g. Niemelä & Tynni 1979, Saarnisto & Salonen 1995, Donner 1995, Nenonen 1995). The Suupohja region consists of the municipalities of Isojoki, Jurva, Karijoki, Kauhajoki, Teuva, Kaskinen, Kristiinankaupunki and Närpiö (Fig. 1).

Former studies have been concentrated almost exclusively on eskers, which usually have distinct geomorphology and where abundant open sections are observable. During the inventory of sand and gravel resources in the Kauhajoki district (Kurkinen et al. 1994), also other types of accumulations with multiple stratigraphy were found, and which do not have the geomorphological form of an esker. In these accumulations different types of till units and both coarse- and fine-grained sorted deposits can be observed superimposing upon each other. The drilling made by the Geological Survey of Finland

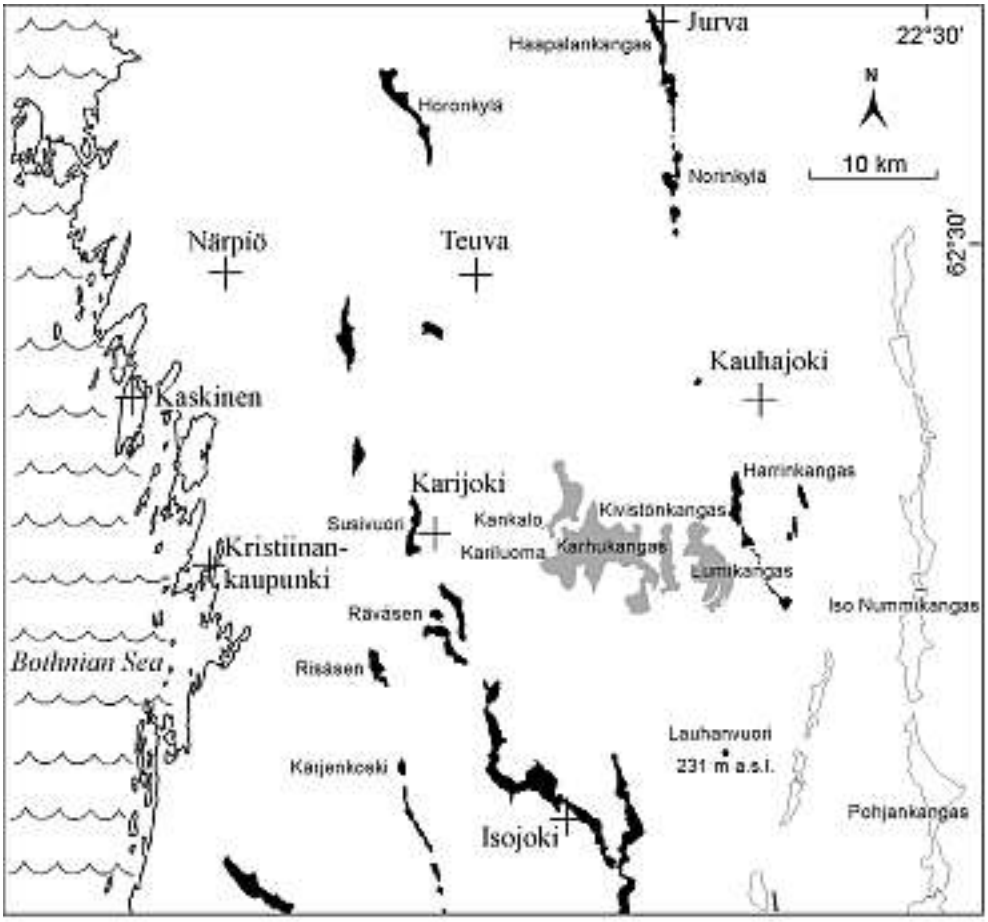


Fig. 1. The Suupohja region with the thick multilayer accumulations (grey areas). The "till-covered eskers" (see Niemelä 1978) are marked as black and eskers without till cover as white. Sites described in previous papers are also indicated.

at Karhukangas penetrated 95.3 m thick Quaternary sediment sequence and ended into sandstone (Huhta 1997). The drill core revealed twelve distinguishable lithofacies, which obviously have been deposited in different sedimentary environments. Up to five glacial-deglacial phases can be constructed according to the drilling, although it is possible that some of the multiple layers have been deposited during a single glaciation due to glacier oscillation or from different parts of a glacier (see e.g. Brodzikowski & Van Loon 1991). Different till layers, glaciofluvial, shore and eolian sediments, permafrost structures and paleosoils refer to two separate glacial and deglacial phases (before the Holocene) interpretable even in the uppermost few metres (Fig. 2). The interpretation of different lithofacies in the drillings, open sections and test pits is based on grain-size composition, texture, color, organic content, sedimentary structures and directional elements (when observable/measureable). In addition, refraction seismic and ground-penetrating radar measurements were utilized (see Pitkäranta 1996, Pitkäranta submitted).

All the observed sedimentary units in the Suupohja region have been grouped to form the Suupohja Group, which consists of six Quaternary formations. The sedimentary units have been described recently in Pitkäranta (submitted) and in previous papers (see e.g. Niemelä & Tynni 1979, Donner 1988, Gibbard et al. 1989, Bouchard et al. 1990, Nenonen 1995, Huhta 1997, Kujansuu & Uutela 1997). Four new and two previous formation name proposals are presented in the following. In addition, formal name for the Paleozoic sandstone (see Simonen & Kouvo 1955, Lehtovaara 1982) is proposed.

Formal lithostratigraphy in glaciated areas

Utilizing formal lithostratigraphy helps in communication. It is a tool when dealing with more or less complex lithostratigraphy, and also clarifies correlation between units in different areas. The presupposition is that the units, which are going to be named, will be duly defined, characterized, described and proposed (see instructions in the International Stratigraphic Guide,

Hedberg 1976 [1st edition] and Salvador 1994 [2nd edition]). Using formal names diminishes confusing, subjective and nonconsistent descriptive terms.

Assigning formal names to lithostratigraphical units has been rarely used in Finland and in Fennoscandia as a whole (cf. Gibbard 1992). Reasons for this are probably lack of tradition to name units and the fact that Quaternary deposits in Fennoscandia are relatively thin and heterogenous, and particular lithofacies associations occur usually in only limited area. This may lead to a jungle of lithostratigraphic names, because both litho- and chronostratigraphic correlation between different areas is difficult, and hence combining units from different areas is often impossible. Deposits from the same time interval and even the same sedimentary event may vary a lot from one place to another. Also sediments originating from a single sedimentary event can vary a lot in a vertical section (see e.g. Eyles et al. 1983, Johnson & Hansel 1990, Brodzikowski & Van Loon 1991). Furthermore, the lack of organic sediments and fossils make the correlation uncertain. If they are present, possible redeposition must be considered. Finally, detailed descriptions of the deposits have concerned only limited areas, and even then the descriptions are heterogenous. They don't necessarily fill the requirements, which should be taken into account when assigning formal names (see Salvador 1994).

One problem in the procedure of formal naming is the designation of stratotypes (type sections). Natural sections, such as river terraces and seaside cliffs, are rare in these areas. Observations have been made usually at artificially made open sections (e.g. quarries, construction sites), where excavation is going on and the original description sites will easily be destroyed. On the other hand, where excavation has ceased, section walls will collapse or they have been filled. Consequently, designated stratotypes lose their information/position as a representative observation site for particular lithostratigraphic units. Thus, in addition to the designation of the original primary stratotype (holostratotype), also supplementary stratotypes (parastratotypes) should be indicated if possible (see Salvador 1994). One possibility is also to establish rules, how to preserve valuable geological sections.

Proposed formal names for the Suupohja Group

Up to seven different till units, three glaciomarine/-lacustrine units, three glaciofluvial, two wave-generated and three eolian sediment units (including Holocene sediments) have been interpreted in the Suupohja region. All the observed lithofacies and their formal names are summarized in Table 1. Generalized cross-section from Karhukangas to Lumikangas with identified lithofacies is illustrated in Figure 2. Five

Table 1. Summary of the lithostratigraphy, brief description of the deposits and tentative chronostratigraphy of the Suupohja Group sediments. The locations of the type sections are indicated in the Finnish co-ordinate system, Zone 1. Elevation (Z) in meters above present sea level.

Lithostratigraphic names	Observable units	Short description of the sediments with interpretation	Chronostratigraphy	Type section or type area
Lumikangas Formation		Stratified sand and gravel, shore deposit	Holocene	X=6910.00 Y=1559.80 Z=165
Kauhajoki Till	Two different types (see Bouchard et al. 1990)	Dark massive basal till, high content of fines (35 – 60 % < 0.0625 mm)	Middle or Late Weichselian	X=6915.50 Y=1560.45 Z=140
Harrinkangas Formation	Several different units (see Gibbard et al. 1989)	Glaciofluvial gravel, sand and silt, compact, deformed, permafrost structures	Late Saalian – Eem	X=6915.50 Y=1560.45 Z=140
Kariluoma Till	At least two different types	Light compact sandy lodgement till, 10 – 30 % < 0.0625 mm	Late Pleistocene (Late Saalian)	X=6909.71 Y=1549.02 Z=140 *
Kankalo Sand	Three units	Stratified, cross-bedded and ripple cross-laminated medium-grained glaciofluvial, shore and eolian sands, compact, deformed, permafrost structures	Middle or Late Pleistocene (Middle Saalian) (?)	X=6914.42 Y=1547.56 Z=125 *
Karhukangas Formation	Nine units (see Huhta 1997)	Mainly tills and glacilacustrine/-marine sediments superimposing each other	Early or Middle Pleistocene (?)	Type area: central part of Karhukangas
Lauhanvuori Sandstone	(see Simonen & Kouvo 1955, Lehtovaara 1982)	Lithified sandstone (minor occurences of conglomerate)	Paleozoic	Type area: Lauhanvuori hill

* Also supplementary type sections designated.

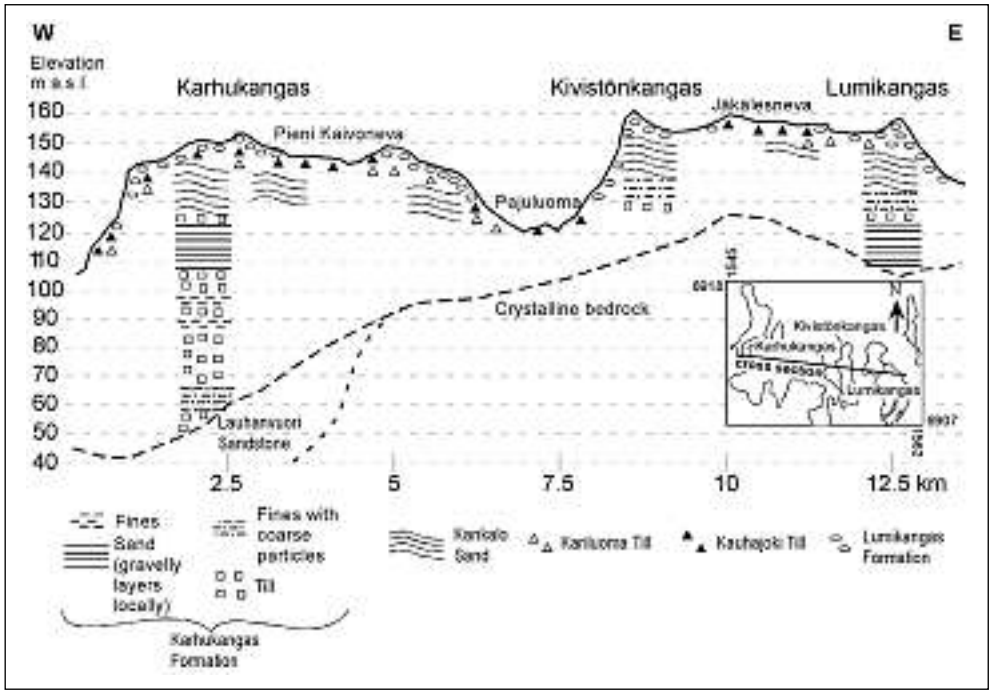


Fig. 2. Generalized cross-section with observed lithostratigraphy through Karhukangas, Kivistönkangas and Lumikangas.

new formation names have been proposed: Lauhanvuori Sandstone, Karhukangas Formation, Kankalo Sand, Kariluoma Till and Lumikangas Formation. In addition, two formations have been described and named earlier: the Harrinkangas Formation and the Kauhajoki Till (Gibbard et al. 1989, Bouchard et al. 1990). Furthermore, the Ostrobothnia Geosol (Kujansuu et al. 1991) has been proposed as the formal name for the paleosoils found in the Suupohja region. All the formations mentioned above are included to the Suupohja Group.

Age estimations of the Suupohja Group sediments

Chronostratigraphy of the Suupohja Group sediments is by far unknown at present (see Table 1). Several thermo- and optically stimulated luminescense datings have been made on the deposits correlative to the Harrinkangas Formation (see Nenonen 1995 with references). They refer to ages ranging from around 160 ka to 70 ka, i.e. from the Late Saalian to the beginning of the Middle Weichselian Substages. Consequently, it can be assumed that all the sediments below the Kauhajoki Till are older than the Middle Weichselian Substage (Table 1 and Fig. 2). Also lithostratigraphy and particular post-depositional features suggest older than the Middle Weichselian age for the sediments below the Kauhajoki Till (see Pitkäranta submitted). Two TL datings made from the Kankalo Sand referred to infinite ages (Kujansuu & Uutela 1997), although there were some uncertainties with the timing. The Kankalo Sand is probably older than the Harrinkangas Formation, and pre-Late Saalian age can be estimated for that formation. The ages of the different lithofacies of the Karhukangas Formation remain unknown, as no datings are available. Middle and even Early Pleistocene age can be estimated for the deeper layers. Microfossil and organic contents in the fine-grained sediments are negligible, and they don't reveal the sedimentation environments.

The paleosoils (Ostrobothnia Geosol) are interpreted to originate from the prolonged ice-free period extending from the Eemian to the Middle Weichselian Substage, i.e. ~ 125 – 74 ka before present (see also Kujansuu et al. 1991, Kujansuu 1992, Hütt et al. 1993). Permafrost structures in the Harrinkangas Formation and the Kankalo Sand indicate fairly long and cold ice-free period in the Early Weichselian, although it is also possible that the permafrost structures in these two formations originate from different cold periods. Both paleosoils and permafrost features are good marker horizons in lithostratigraphic and chronostratigraphic correlation.

Conclusions

The Suupohja Group consists of six Quaternary formations. This division is based on the interpretation that the named formations have been deposited

in different sedimentary environments and probably in different glacial-deglacial phases. The lowermost Karhukangas Formation has been observed only in drillings and indirectly in refraction seismic and ground-penetrating radar measurements. Tills and glaciolacustrine or glaciomarine deposits predominate in that formation. The upper parts of the Kankalo Sand, as also the other formations above it, can be detected at open sections. They are mainly tills and glaciofluvial deposits. Relatively thin layers of shore and eolian deposits have been determined in the upper parts, too. The Kankalo Sand, Kariluoma Till, Harrinkangas Formation, Kauhajoki Till and Lumikangas Formation are clearly distinguishable units and probably of different origin, and hence they have been separated into different formations. The majority of the Suupohja Group sediments have been deposited in glacial or near-glacial environments.

Fine-grained waterlain sediments occur in the deeper parts of the deposits, because during the time of their deposition, the area was relatively low-lying terrain. Shallow water, glacial and terrestrial sediments predominate in the upper parts.

Because there is a rich diversity of Quaternary stratigraphy in the Suupohja region, it is particularly suitable area for formal lithostratigraphical examination.

REFERENCES

Bouchard, M.A., Gibbard, P. & Salonen, V.-P. 1990. Lithostratotypes for Weichselian and pre-Weichselian sediments in southern and western Finland. Bulletin of the Geological Society of Finland 62, 79–95.

Brodzikowski, K. & Van Loon, A.J. 1991. Glacigenic sediments. Developments in sedimentology 49. Amsterdam: Elsevier, 674 p.

Donner, J. 1988. The Eemian site of Norinkylä compared with other interglacial and interstadial sites in Ostrobothnia, western Finland. Annales Acadamicae Scientiarum Fennicae A III 149, 31 p.

Donner, J. 1995. The Quaternary History of Scandinavia. World and Regional Geology 7. Cambridge: Cambridge University Press, 200 p.

Eyles, N., Eyles, C.H. & Miall, A.D. 1983. Lithofacies types and vertical profile models; an alternative approach to the description and environmental interpretation of glacial diamicton and diamictonite sequences. Sedimentology 30, 393–410.

Gibbard, P.L. 1992. Formal stratigraphy in the Pleistocene of Finland. Bulletin of the Geological Society of Finland 64, 125–132.

Gibbard, P., Forman, S., Salomaa, R., Alhonen, P., Jungner, H., Peglar, S., Suksi, J. & Vuorinen, A. 1989. Late Pleistocene stratigraphy at Harrinkangas, Kauhajoki, western Finland. Annales Academiae Scientiarum Fennicae A III 150, 36 p.

Hedberg, H.D. (ed.) 1976. International Stratigraphic Guide – A guide to stratigraphic classification, terminology and procedure. New York: John Wiley and Sons, 200 p.

Huhta, P. 1997. Almost 100 m of Quaternary deposits on sandstone at Karhukangas, Kauhajoki, western Finland. In: Autio, S. (ed.), Geological Survey of Finland, Current Research 1995–1996. Geological Survey of Finland, Special Paper 23, 89–91.

Hütt, G., Jungner, H., Kujansuu, R. & Saarnisto, M. 1993. OSL- and TL-dating of buried podzol and overlying sands in Ostrobothnia, western Finland. Journal of Quaternary Science 8, 125–132.

Johnson, W.H. & Hansel, A.K. 1990. Multiple Wisconsinan glacigenic sequences at Wedron, Illinois. Journal of Sedimentary Petrology 60, 26–41.

Kujansuu, R. 1992. Paleosols as Quaternary stratigraphical key horizons in Ostrobothnia, western Finland. Bulletin of the Geological Society of Finland 64, 161–167.

Kujansuu, R., Saarnisto, M., Räisänen, M.-L. & Hansel, A. 1991. Fossil soil of Kärjenkoski and its correlatives in Ostrobothnia, western Finland. In: Autio S. (ed.), Geological Survey of Finland, Current Research 1989 – 1990. Geological Survey of Finland, Special Paper 12, 119–126.

Kujansuu, R. & Uutela, A. 1997. Palaeozoic acritarchs in till-covered sand deposits at Kauhajoki, Western Finland. In: Autio S. (ed.), Geological Survey of Finland, Current Research 1995 – 1996. Geological Survey of Finland, Special Paper 23, 93–98.

Kurkinen, I., Ristaniemi, O. & Pitkäranta, R. 1994. Kauhajoen kunnan sora- ja hiekkavarat. Vaasan läänin seutukaavaliitto, Sarja D:33, 56 p. [in Finnish]

Lehtovaara, J. 1982. Palaeozoic sedimentary rocks in Finland. Annales Academiae Scientiarum Fennicae, A III 133, 35 p.

Nenonen, K. 1995. Pleistocene stratigraphy and reference sections in southern and western Finland. Geological Survey of Finland, Miscellaneous Publications, Ph.D. Thesis, 94 p.

Niemelä, J. 1978. Etelä-Pohjanmaan sora- ja hiekkamuodostumien geologinen tausta (Summary: The geological background of the sand and gravel deposits of southern Ostrobothnia). Papers of the Engineering-Geological Society of Finland 12, 15 p. [in Finnish]

Niemelä, J. & Tynni, R. 1979. Interglacial and interstadial sediments in the Pohjanmaa region, Finland. Geological Survey of Finland, Bulletin 302, 48 p.

Pitkäranta, R. 1996. Moreenipeitteiset lajittuneet kerrostumat Kauhajoen alueella ja eri tutkimusmenetelmien soveltuvuus niiden tutkimiseen (Summary: Till-covered sorted deposits in the Kauhajoki area, southern Ostrobothnia and the suitability of investigation methods to explore them). University of Turku, Department of Geology, Unpublished Ph.L. Thesis, 121 p. [in Finnish]

Pitkäranta, R. Lithostratigraphy of the pre-Weichselian extensive erosion remnants in the Suupohja region, western Finland. Boreas, submitted.

Saarnisto, M. & Salonen, V.-P. 1995. Glacial history of Finland. In: Ehlers, J., Kozarski, S. & Gibbard, P. (eds.), Glacial Deposits in North-East Europe. Rotterdam: A.A.Balkema, 3–10.

Salvador, A. (ed.) 1994. International Stratigraphic Guide – A guide to stratigraphic classification, terminology, and procedure. Second Edition. Trondheim: the International Union of Geological Sciences and The Geological Society of America, 214 p.

Simonen, A. & Kouvo, O. 1955. Sandstones in Finland. Bulletin de la Commission Géologique de Finlande 168, 57–86.

Quaternary studies in the northern and Arctic regions of Finland
Edited by Antti E.K. Ojala
Geological Survey of Finland, Special Paper 40, 97–111, 2005.

THE UNIQUE MORAINE MORPHOLOGY, STRATOTYPES AND ONGOING GEOLOGICAL PROCESSES AT THE KVARKEN ARCHIPELAGO ON THE LAND UPLIFT AREA IN THE WESTERN COAST OF FINLAND

by
Olli Breilin[1)], Aarno Kotilainen[1)], Keijo Nenonen[1)], Matti Räsänen[2)]

[1)]*Geological Survey of Finland, P.O. Box 96, FIN-02151, Espoo, FINLAND*
[2)]*Department of Geology, FIN-20014, University of Turku, FINLAND*
E-mail: olli.breilin@gtk.fi, aarno.kotilainen@gtk.fi, keijo.nenonen@gtk.fi, matti.rasanen@utu.fi

European Union

Breilin, Olli, Kotilainen, Aarno, Nenonen, Keijo & Räsänen, Matti, 2005. The unique moraine morphology, stratotypes and ongoing geological processes at the Kvarken Archipelago on the land uplift area in the Western coast of Finland. *Geological Survey of Finland, Special Paper 40*, 97–111, 18 figures.

The Kvarken Archipelago is situated in the centre of the Fennoscandian glacioisostatic land uplift area, with an overall net uplift rate of 8 to 8.5 mm per year. At a maintained uplift rate Finland and Sweden will become connected with a land bridge across the Kvarken strait in about 2500 years. The Bothnian Bay will then become the largest freshwater lake in Europe.

The rapid land uplift gains approximately 100 hectares of new land emerging from the Baltic Sea annually at the Kvarken archipelago, Western low topography coast of Finland. The Kvarken Archipelago is characterized by extensive moraine ridge topography and a shallow brackish sea (= low salinity 4–5 per mil). The area includes approximately 7000 islands and islets and a total shoreline of approx. 3000 kilometres. The bedrock is eroded to a peneplain already during the late Proterozoic thus forming unique platform for the study of rapid isostatic land uplift and its effects on geological and coastal processes as well as the biological successions of plant communities.

The Quaternary deposits on top of crystalline bedrock is composed of moraine formations, dating back to the last deglaciation. The major geomorphologic feature, which makes the Kvarken Archipelago area extraordinary, is the spectacular De Geer moraine fields, deposited during the gradual deglaciation of the continental ice sheet. The De Geer moraines are exceptionally abundant, well formed and representative and frequently appear in large fields within the area. Also, hummocky moraines and other types of transversal moraine ridges occur.

At the modern sea bottom in the Kvarken Strait, where the moraines have not yet been exposed to wave and current activity, the moraines have their original form created by the inland-ice. Owing to the ongoing land uplift process these are rising above the sea surface, as further invaluable geological records. Finland has made application of The Kvarken Archipelago to UNESCO`s World Heritage Committee on January 2005 to be nominated as a World Heritage object on Natural criterion (i):"be outstanding examples representing major stages of the earth's history, including the record of life, significant ongoing geological processes in the development of landforms, or significant geomorphic or physiographic features".

Key words (GeoRef Thesaurus, AGI): archipelagos, geomorphology, glacial features, moraines, De Geer moraines, hummocky moraines, stratotypes, uplifts, marine geology, Quaternary, Merenkurkku, Kvarken, Finland

Introduction

The Kvarken Archipelago has undergone a long sequence of geological processes. What we can see today is one part of the geologic evolution of the earth's crust: crystalline bedrock from ancient times and overburden representing young geologic processes as a heritage of the ice age. Ongoing geologic processes are rapidly and invariably changing the face of this unique area over the course of a human's lifetime.

The glacial events and formations of the Quaternary ice age have built up the unique landscape and landforms of the Kvarken Archipelago (Fig. 2). The long lasting erosion and peneplanation of the Precambrian bedrock form a peculiar platform for the dynamic, ongoing geological processes. Since the early days of glacial geology, the Kvarken area has been the focus of scientific research dealing with postglacial land uplift, moraine morphology and deglaciation.

The Northern Kvarken is the narrowest part of the Gulf of Bothnia in the northern Baltic Sea, connecting Finland's Ostrobothnia (Pohjanmaa) and Sweden's Västerbotten. The distance from the Finnish to the Swedish coast is 80 km, with only 25 km between the outermost islands (Fig. 1). Northern Kvarken also forms a submarine sill, which separates the Bothnian Sea in the south from the Bothnian Bay in the north. The Kvarken Archipelago is situated at the eastern part of the Northern Kvarken. There are more islands than at any other archipelago in the Gulf of Bothnia. However, only few islands are located along the Swedish coast.

The bathymetry of this area has changed considerably since the last deglaciation, mainly due to the relatively strong land uplift. At present, the Northern Kvarken is very shallow (0–25 m) and shoaly, but during and just after the last deglaciation (around 10 000 years ago) it was submerged 250 – 280 meters. The area includes 6,550 islands and a total shoreline of 2,840 kilometres, a large number of peninsulas and bays, and extensive stony seashores in the Kvarken Archipelago area. The land increase in the area is about 100 ha/year based on calculations from 1:20 000 scale digital base maps of different ages. The highest hill peak (17.5 m a.s.l.) is situated on Replot Island.

During the last glaciation, the Northern Kvarken was located close to the centre of the Weichselian ice sheet, which reached ~2800 m thickness during the glacial maximum (Svendsen et al 2004). The islands of the region are smooth, glacially eroded, low rocky islands, which are characterized by till boulders on the shores. Also typical for the very special landscape of the Kvarken area are boulder-rich ridges so-called De Geer moraines. On the Swedish side of the Northern Kvarken, the shoreline is steeper and the archipelago smaller with just a few islands. The landscape in

Fig. 1. Location of the proposed world heritage area (Ollqvist & Rinkineva-Kantola 2004).

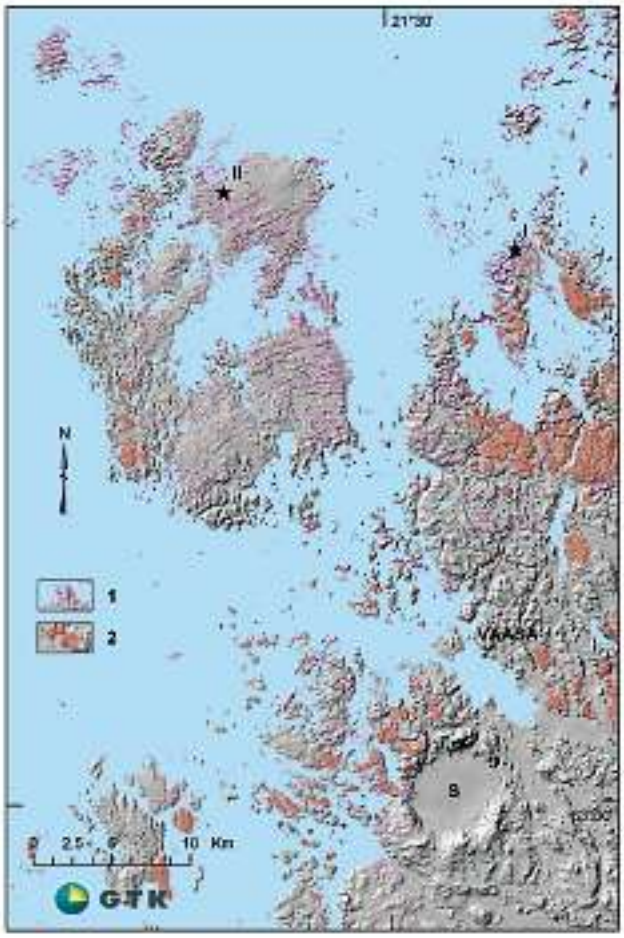


Fig. 2. The diverging of the de Geer moraine and large transversal moraine ridge clusters in the north shown by geographical elevation model. The Söderfjärden meteorite impact crater (S) is shown near Vaasa in the southern part of the image. 1 = de Geer moraine and large transversal moraine ridge, 2 = hummocky moraine. Star dots I and II = locations of excavations on moraine ridges (see Figs. 6 and 7) (Compilation: M. Paalijärvi).

Swedish side is dominated by streamlined till ridges, drumlins and flutings and elevated shore deposits, shingle field on shore slopes though De Geer moraines occur there too but more sparsely.

The World Heritage application of The Kvarken Archipelago is stated by arguments of outstanding universal value for the understanding of how glacial and deglaciation processes form a landscape. The Kvarken Archipelago is also the most representative area in the world for studying moraine archipelagos and the land uplift phenomena. The High Coast World Heritage and the Kvarken Archipelago represent complementary examples of post-glacial uplifting landscapes (Ollqvist & Rinkineva-Kantola 2004, Breilin et al. 2004) (Fig. 1).

As a complement to the High Coast World Heritage Site, The Kvarken Archipelago is proposed to the World Heritage List based on Natural criterion (i):"be outstanding examples representing major stages of the earth's history, including the record of life, significant ongoing geological processes in the development of landforms, or significant geomorphic or physiographic features" (Ollqvist & Rinkineva-Kantola 2004, Breilin et al. 2004).

Glacial formations

The most widespread surficial and seafloor sedimentary deposit in the Kvarken area is glacial till and its moraine formations. Till consists of varying amounts of boulders, stones, gravel, sand, silt and

GEOLOGICAL FEATURES OF THE KVARKEN ARCHIPELAGO

QUATERNARY DEPOSITS
- Quaternary deposits consist of young mainly glacial sediments approx. 13 000 – 10 000 BP years old. Maximum thickness up to 50 m, average 5 m.
- Complex deglaciation pattern with several ongoing ice-flow stages.
- Main soil types are glacial till, boulder fields and marine/littoral sediments and young organic deposits like gyttja and peat bogs.
- Typical glacial formations are De Geer moraines and large traversal "Rogen type" moraines, which form large moraine fields with hundreds of formations. Also, some hummocky moraine fields and minor drumlins occur.
- Rapid glacial isostatic land uplift (8 – 8.5 mm/year) and new land emerges from under the sea with an area of 100 ha/year.

BEDROCK
- The hard crystalline bedrock belongs to the Precambrian Svecofennian schist belt.
- The bedrock consists mainly of gneiss, diatexite, granites and diabase.
- The history of bedrock contains 11 different phases, aging from 1,9 Ga to 520 Ma.
- Palaeozoic sediments began to deposit on the eroded Precambrian peneplain 520 million years ago.
- The main part of these sediments is also missing (due to peneplanation erosion processes).

Fig. 3. Schematic figure of the moraine formations in the Kvarken area. 1. Drumlins, 2. Flutings, 3. Large transversal moraines (Rogen type), 4. De Geer moraines, 5. Hummocky moraines, 6. Boulder-rich surface, 7. End moraines, 8. Latest ice flow direction (Drawing: H. Kutvonen).

clay. Till generally lies directly on the bedrock and largely follows its surface configuration. Till also commonly forms surficial configurations like various moraine ridges and hummocks (Fig. 3). Numerous De Geer moraines are a specific feature of the Kvarken Archipelago. The best area to study De Geer moraines is in the Björkö Svedjehamn area (Fig. 4).

Moraines parallel to the ice direction

A drumlin is the oval-shaped streamlined ridge formed beneath an ice-sheet as it moved over the terrain. Drumlins and drumlinoid formations are often found in groups and the ridges may extend for several kilometers. Drumlins are composed mainly of basal till or lodgement till. Fluting ridges are either depositional or erosional small glacial features on the basal till surfaces, indicating the last ice flow direction.

Both drumlins and flutings form a small field at Norra Vallgrund where the younger transversal moraines cover the streamlined parallel moraine forms. In the Swedish side of the Kvarken both drumlins and flutings are the major moraine forms on till areas.

Fig. 4. Helicopter view of the De Geer Moraines. Björkö Svedjehamn. (Photo: A. Hämäläinen, 2000)

Moraines parallel to the ice margin

End moraines belong to this class and may either be large or small, short or long. The end moraines were formed along the ice margin and often have an asymmetrical shape with a gentle stoss-side and a steeper lee-side. A closely related type of moraine, the De Geer moraine, occurs in clusters in lowland areas (Fig. 4). Earlier, it was thought that the moraines were formed at the ice margin and were a type of end moraine. De Geer moraines are most commonly till ridges up to 5 m high, from 10–50 m wide and a couple of hundred meters long. In rare instances, the height exceeds 5 m and length 1000 m. The moraines occur at 40 – 300 m intervals in large groups, mostly in low-lying landscape areas. The Replot and Björkö areas in the Kvarken Archipelago are the best examples of clusters of De Geer moraines in Finland (Aartolahti 1988).

In the Kvarken Archipelago area, the amount of De Geer moraine ridges is greatest and they occur in very compact formation swarms. The width of the formations is often over 50 m and the formations are symmetrical (Laaksonen 1994). In some areas, the formations seem to be related or deposited on top of drumlin forms and other hummock moraine forms like large transversal Rogen type ridges. In the Märaskär and Köklot areas, De Geer moraines are deposited on top of larger Rogen type moraine formations and the long axis of both moraine formations forms a network pattern in the terrain and archipelago. Laaksonen (1994) put forward a competing hypothesis of stabile laminar basal flow of ice as an explanation of the wave like pattern and symmetric forms of the De Geer moraines and swarms. It is evident that both De Geer moraines, transverse basal till ridges (Rogen type moraines or ribbed ridges) and radial streamlined moraine ridges (drumlins) occur in the same area of Replot and Björkö islands. There may also be a connection between the clusters of De Geer moraines and tectonic features (Lundqvist 2000).

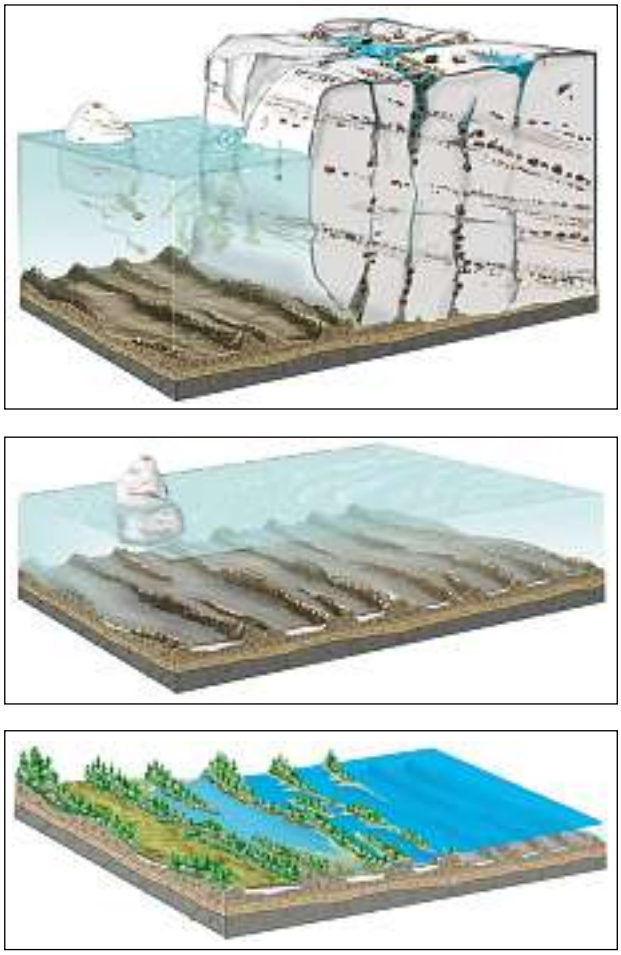


Fig. 5. The suggested formation of De Geer moraines (Drawings: Harri Kutvonen).

The De Geer moraine formation were first described in Sweden by De Geer (1889) and called De Geer moraines by Hoppe (1957) or washboard moraines by Mawdsley (1936). In the case of the Kvarken Archipelago, the water depth during deglaciation and formation of the moraine ridges was 250–280 m. According to the current moraine genesis explanation, the moraine ridges were formed in the crevasses running parallel to the ice margin in sub -aquatic conditions. Huge icebergs calved at the ice front and the De Geer moraine reflects the probable position of the retreating ice margin (Fig. 5).

Hummocky moraines

Hummocky moraines principally occur in valleys and in flat-lying areas. Hummocky moraines are irregular and non-oriented formations, usually 5–20 m high, and form a mosaic of lakes, tarns, and peatlands. The moraines were deposited in the Kvarken area beneath the melting and thinning glacier front in sub-aquatic conditions. Material in these formations is usually washed and coarse grained till (melt out till). Most of the hummocky moraine formations were deposited in the final stage of the last deglaciation. In a way, they witness the non-oriented deposition character of till material during the formation of De Geer moraines. In some localities, the texture of the faulted and fissured glacier front can be seen in the hummocky relief, sometimes resembling ring-ridge type of deposition like in Pulju moraines or ribbed moraines in Northern Finland (comp. Sarala 2003, 2004 and Johansson & Nenonen 1991). Some hummocks are just heaps of till that were deposited, squeezed or that flowed in a crevasse or cavity beneath the melting glacier.

A Rogen moraine is a type of hummocky moraine or ribbed moraine characterized by ridges, that are irregular in detail but largely at right angles to the

direction in which the ice was moving (comp. Lundqvist 1989, Hätterstrand 1997 and Sarala 2003, 2004). Rogen type moraines are often composed of basal till or lodgement till and deposited clearly in sub glacial conditions probably in the same zones as the drumlins. Transversal basal till ridges occurs in swarms and often inside drumlin fields. Some Rogen type moraine formations in the Kvarken area have developed elongated tails parallel to the last ice flow directions, thus showing a relationship to the drumlin forming processes. Rogen-like transversal moraine formations occur in the Köklot, Mickelsörarna and Valsörarna areas.

The drumlins and large transversal moraine ridges (Rogen type) were formed below the ice sheet, some 200 – 700 km inside ice margin. At that time, huge ice lobes filled the Bothnian Bay area and the ice flow was roughly south-southeast as shown by striations and drumlin orientations on maps (comp. Bargel et al. 1999). The term hummocky moraine is used to describe all kinds of moraine hummocks. Cuts in the moraines and documentations of the sections give tools for a more detailed classification. In recent investigations, new evidence of complex ice flow directions have been discovered (Geonat 2004). Therefore, the Kvarken Archipelago might become one of the key areas for understanding the Early- and Mid- phases of Weichselian glaciation.

One striking phenomenon in the Kvarken Archipelago is the boulder rich till surfaces even in shallow sea areas, like in the Ikmo Lillön area, Halsön Island and the Bergö Gaddarna rocky islets. Some of the boulders are huge erratics transported by flowing glacier ice or floating icebergs in the Ancylus Lake. The granitic rock types are susceptible to intensive cubic cracking and thus large boulders and erratics are easily carried by the glacier from rock outcrops and possible preglacial inselberg or tor like bedrock formations. This boulder rich undulating moraine terrain is an example of glacioaquatic and subaquatic till deposition and represents the youngest till deposit and moraines in the area.

Structure of the De Geer moraine and large transversal moraine ridge at Björköby and Köklot

Three representative moraine formations in the GEONAT research area were chosen for further studies. The Björkö Skagback De Geer moraine formation represents a well-shaped prominent higher moraine ridge with a smooth proximal side and steeply dipping distal side. Two test pits and one trench were excavated in this formation. The Björkö Ohls De Geer moraine formation represents a lower moraine ridge were the proximal and distal sides are both smoothly dipping (location II in Fig. 2). One study trench was excavated in this formation. Both De Geer ridges are located within the Björköby De Geer moraine field where the terrain is covered with numerous higher and lower moraine ridges at less than 100- meter intervals. The third chosen formation at Köklot Furuskäret represents larger, more prominent transversal moraine formations of Rogen type (location I in Fig. 2). De Geer moraine formations are on a transversal position and overlie the larger Köklot moraines especially north from Köklot in the Mickelsörarna area.

Three test pits were excavated in the formation at Köklot. The cuttings and trenches in the Björkö Skagback and Björkö Ohls De Geer moraine formations revealed that both lodgement till and melt out till are present in the formations. The till fabric in

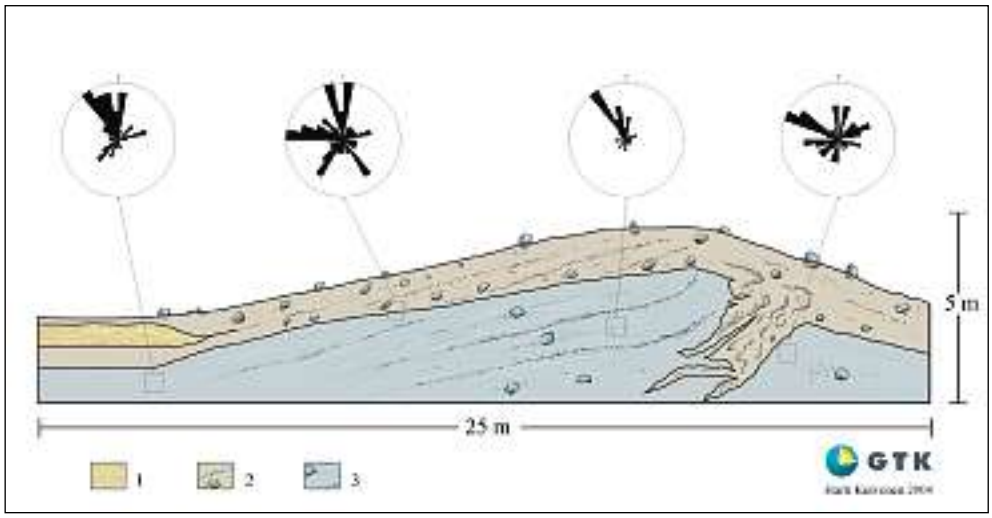


Fig. 6. Schematic cross section from test pit excavated in the low-lying De Geer moraine Björkö Ohls. Ice flow direction from left to right in drawing. 1= sandy loam, 2= melt out till with deformation structures and 3= basal till with some deformation structure (location II in Figure 2, Drawing: H. Kutvonen)

both moraine types showed relatively good preferred orientation along the latest ice flow direction though some fabric analyses also showed transversal peaks or were not oriented (Fig. 6).

The structure of the till material shows that shearing and deformation was present during the depositional process. Shear planes, glaciotectonic folds, thrust structures and good preferred orientation in the till fabric shows the presence of actively flowing ice and deposition in a subglacial environment. The texture of the till material is a matrix supported in the lodgement type till and both the matrix and clast were supported in the melt out till type, which also shows sorting of sand and gravel material. The sorting and coarser texture of the melt out till type demonstrates the presence of water in the depositional process. The petrographic composition of the till types in the Björkö formations shows quite similar transport conditions in both formations and moraines and results are typical of subglacial basal tills.

Results of the excavations of the larger Köklot Furuskäret moraine formation differ considerably from the Björkö formations. The till type in the Köklot formation is of a melt out type. The structure of the till shows sorting, sandy wrappings under and over stones and boulders, abundantly sorted sandy and gravelly lenses, bed and layer structure in till and some folding and bending of the sorted layers (Fig. 7). The till fabric is poorly oriented. One fabric analysis shows a northerly fabric, which coincides with younger northerly striations and De Geer moraines deposited by a northeastern ice flow. The moraine in the Köklot formation is over consolidated and was clearly deposited in subglacial conditions. The petrographic composition differs clearly from the Björköby area showing abundant granites and porfyric granodirites in the ice flow direction.

The studied moraine formations in Björkö and Köklot are clearly of subglacial and subaquatic origin (comp. Benn & Evans 1998). The Björkö De Geer moraine formations shows actively flowing, deforming and pushing behavior of the ice edge rather than the melting, loading and convoluting, crevasse fill phenomena of subglacial ice crevasses (Fig. 8). The depositional environment indicates sub marginal formation of parallel moraine ridges at a calving ice margin (comp. Aartolahti 1995 and Linden et al. 2004). The Köklot large transversal moraine formations show abundant subglacial melting, glacial loading, consolidating of melt out till and probably reactivation of the glacier sole. The Köklot formations probably have a complex origin and there are still riddles to be solved

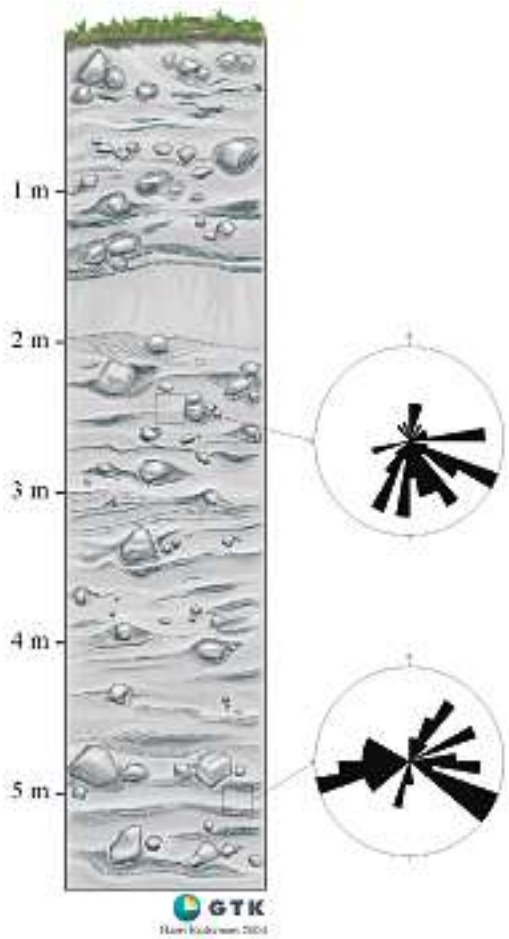


Fig. 7. Schematic profile and photograph from the test pit excavated in large transversal moraine ridge in Köklot. The length of the red knife grip is 10 cm (location II in Fig. 2) (Drawing: H. Kutvonen). (Photo: J. Ojalainen, 2003).

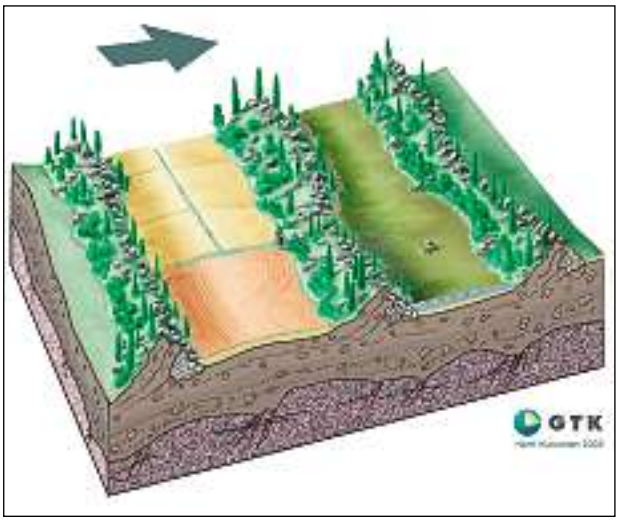


Fig. 8. Schematic picture of the De Geer moraines in the Björkö area. Arrow indicates the last ice flow direction, which is perpendicular to the De Geer moraine ridges. Wedge-like brown basal moraine in the formations illustrates shearing and pushing behaviour of ice edge. Gray moraine inclusions illustrate the melt out facies of the basal till at the distal side of moraine formations (Drawing: H. Kutvonen).

in these formations (comp. Benn & Evans 1998 and Aartolahti 1995).

Svedjehamn gyttja

Excavations at Svedjehamn, Björköby in the northern part of the archipelago discovered a new till covered organic deposit under a 3 – 4 meter thick till cover (Matti Räsänen, personal communication) (Fig. 9). A test pit was excavated at the De Geer moraine's distal site. Later also several sediment core drillings where made to the spot. The till cover over the organic deposit is composed of a sandy till of melt out type. After preliminary laboratory investigations, the organic deposit is interpreted to be gyttja deposited in fresh water conditions. Pollen flora is mainly *Betula*, representing interstadial conditions of Early- or Mid-Weichselian stages (Matti Räsänen, personal communication). The gyttja is just a few meters above the present sea level and is in the lowest topographic position compared with any other till covered organic deposit in Ostrobothnia (Nenonen 1995). This deposit is clear evidence of a much lower sea level stage in the Bothnian Bay area before the last glacial maximum (see Fig. 10). Drillings and earth penetrating radar surveys where made in the location during fall 2004. A detailed study of the biostratigraphy and litostratigraphy of Svedjehamn Gyttja is under work in the University of Turku.

Fig. 9. Svedjehamn gyttja site and example of the gyttja sample. Blue areas in sample shows vivianite – mineral, which is typical for till covered organic deposits. (Photo: J. Ojalainen, 2004)

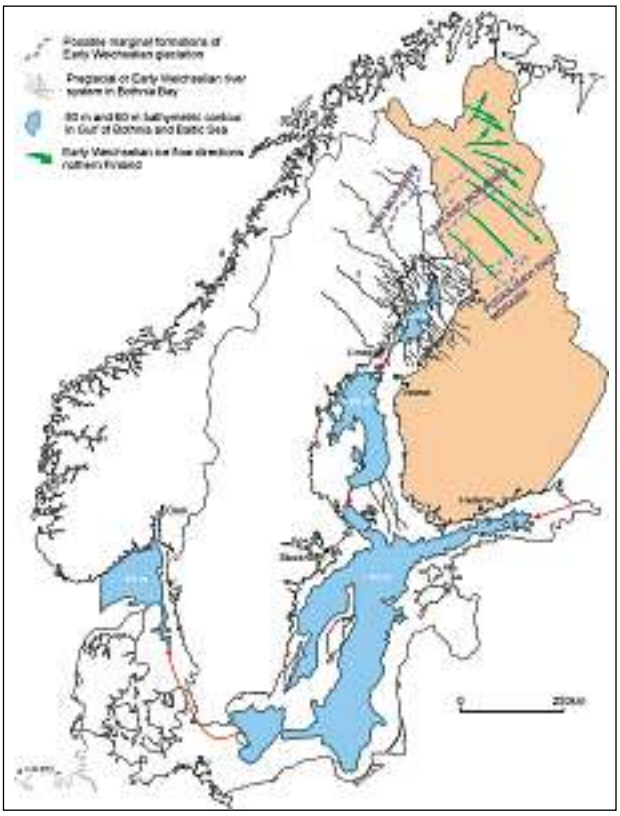


Fig. 10. Bothnian Bay and Baltic Sea area during Weichselian interstadials (Nenonen 1995).

Complex Ice flow pattern in the Kvarken Archipelago

Based on field observations, the phenomena of weak glacial erosion and complex ice flow pattern in the Kvarken Archipelago are strong (Geonat 2004). Evidence of these phenomena is formed in different phases of glaciations. Part of striae and the Svedjehamn organic layer could be of Mid or Early Weichselian age. Field observations of several boulder fields show that rock types are local and in many cases boulder fields are interpreted as erosional remains. This supports the theory of weak glacial erosion in the Kvarken area.

Glacial striaes

On fresh outcrops, glacial striaes are clearly seen in shore areas of thousands of islands and islets (Fig. 11). Faceted, polished surface also appear in some outcrops. Based on field observations of these glacial striae (Fig. 12), ice flow direction has been between 330° – 80° degrees. In the same outcrops, the angle of different directions could be nearly 90° degrees. Also, a direction near 360° degrees has been observed. In some outcrops, the youngest and oldest directions are not same through the area (Geonat 2003, 2004). In some outcrops, the youngest direction is 40° – 65° and in some outcrops 335° – 360°.

Geomorphology

De Geer moraines and large transversal moraine ridges are the main formations of the Quaternary deposits in the Kvarken Archipelago. The direction of the longitudinal axis of moraine formations is complex, which supports the complex glacial striae directions (Figs. 2 and 8). Moraines overlying each other can be seen in some areas. The angle of overlying formations could be up to 90° degrees in northeast parts of the Mickelsörarna area of the archipelago. On the Swedish side, the deglaciation morphology is different. The main feature is a large and intensive drumlin field, which continues from the mainland to the sea bottom (Geonat 2003). The morphology of the sea bottom changes rapidly from drumlin fields to De Geer and large transversal moraine fields approximately 5 km east of Holmö Island on the Swedish coast.

Fig. 11. Crossed glacial striaes (350° – 65°) on a bedrock outcrop on Storskäret Island. Direction of the GPS receiver is 65°. (Photo: O. Breilin, 2003)

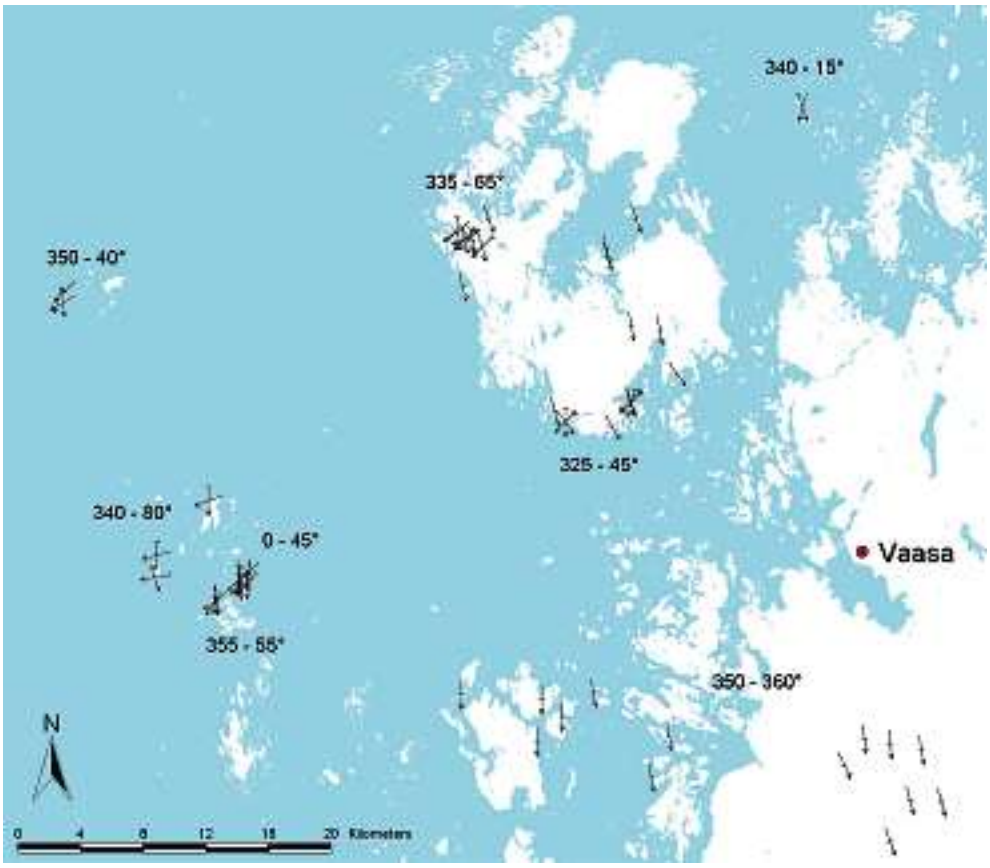


Fig. 12. Location and direction of glacial striations at Kvarken Archipelago (Compilation: J. Ojalainen, Geonat 2003).

Land uplift – historic, present and future

Land uplift studies have a long and respectable history in Finland and Sweden. Changes in shoreline during a human lifetime can be easily observed and had been noticed early. The writer Zachris TOPELIUS depicted late 19th Century the land uplift thus:

"Most noticeable are the effects of this, partly still unexplained, phenomenon. The land rises from the sea, the sea flees, shores are exposed, and the slope is moving forward. Where in days of old ships were sailing, now hardly a ship can travel; where once the fisherman cast his net, now his cows go grazing on the coastal meadow. Banks and rocks appear out of the water, of which no sea chart has had knowledge before; banks expand into islets, these grow together and connect with the mainland. Beaches expand; harbours dry up, seaports must move after the fleeing sea. Every generation of men, new arable land rises from the sea, every century grants Finland a kingdom" (Edlund 1893).

One encounters many a re-owned scientist's name in the field of land uplift research, like the Swede Gerard DE GEER who proved land uplift to be a residual rebound phenomenon from the ice age Finnish geologist Wilhelm RAMSAY, who separated conceptually the isostatic land uplift and the eustatic change of sea level from each other (The Finnish Geodetic Institute, http://www.fgi.fi/yleis/historia_eng.html).

The total depression of the earth surfaces is calculated at 800-1000 m (Taipale & Saarnisto 1990, Eriksson & Henkel 1994, Kakkuri & Virkki 2004). It is assumed that the land uplift will continue some 10 000 to 12 500 years in the Kvarken area and it will still probably result in 100-125 m of isostatic land uplift (Kakkuri 1991). So the uplift will continue until the depression of the geoid is reversed or the next oncoming glaciation begins to load and submerge the Earth's crust in the Kvarken area. The sub-aquatic Kvarken area is a shallow sill with a maximum water depth of 30 m. According to the shore level displacement curve, the sill will be just a strip of water in about 2 500 years (Figs. 13 and 14).

The land uplift had already begun during the melting and thinning of ice 15 000 years ago during the glacial retreat from the Baltic area. During the first

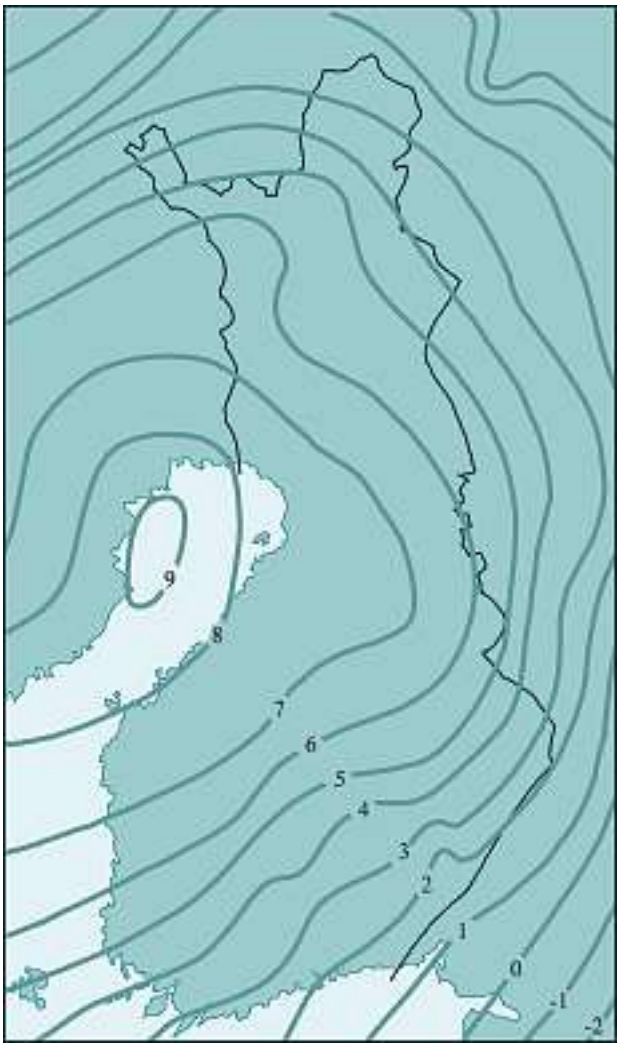


Fig. 13. Present isobases of land uplift (mm per year) in Eastern and Central Scandinavia (Alalammi 1990).

thousand years, the land uplift in the deglaciated areas was calculated to be up to 10 m in 100 years or 100mm/ year (Saarnisto 1981). According to the latest Weichselian ice sheet LGM models, the maximum thickness of the ice had been approximately 2 800 m (Svendsen et al. 2004).

The isostatic land uplift creates not only new land but also many practical problems for the northern Kvarken area. All old harbours are, at present, dry land. The Vaasa-Korsholm harbour that was founded in 12th century is situated 10 km inland from the present Vaskiluoto harbour that was founded in 1890. New land emerges from the shallow sea at a rate of several hectares for individual villages per year. For example, in Replot and Björkö villages 35 hectares of land emerges annually (Palomäki 1988). In the proposed Kvarken Archipelago world heritage area new land emerges from the sea approximately 100 ha annually (Geonat 2004, Ollqvist & Rinkineva-Kantola

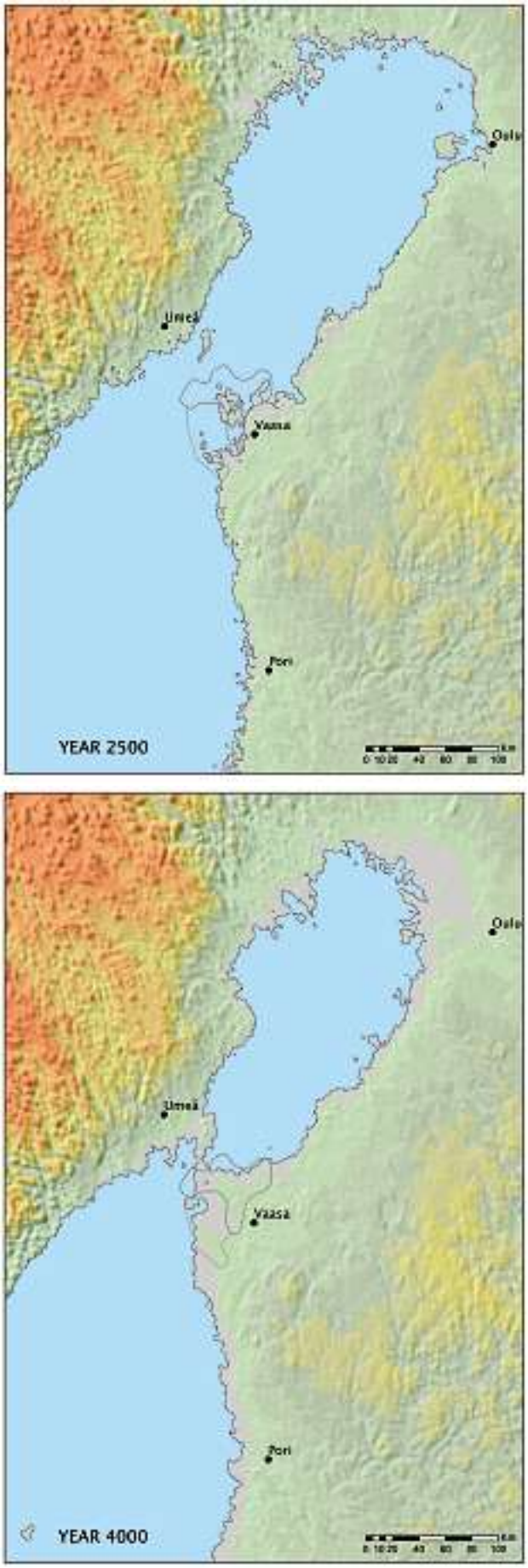


Fig. 14. Future of the Bothnian Bay which is the northern part of the Baltic Sea (DEM Seifert et al. 2001, processing: H. Virkki).

2004). It is estimated that the total new land gain is approximately 7 square kilometres along the Finnish coast (Kakkuri & Virkki 2004).

Also, the construction of deeper harbour basins and canals is a continuous struggle against sea and marine routes that are becoming shallower. A typical phenomenon is summer cottages and their boat shelters lying far inland from the present day shallow shoreline of many low lying islands and peninsulas along the Finnish side of the Kvarken area (see Osala 1988). The land virtually rises from the shallow sea during the span of a human's lifetime. At first, some elongated boulder rich ridges and reefs emerge and seabirds begin nesting there, then the moraine ridges grow together to form elongated bushy moraine islands and finally close into small lagoons. The new soil is fertile and plant cover appears to become established on the shores almost immediately. As land uplift continues, the lagoons become separated from the sea and develop as freshwater ponds or "fladas" and lakes that occasionally get saline intrusions of flooding water from the sea during stormy days. As the vegetation occupies the freshwater ponds wetland development begins, which continues to form raised peatlands in a span of thousand years.

The current relative uplift is about 8.0 mm on the Finnish side of Kvarken area and about 8.5 mm on the Swedish side according to the current postglacial land uplift information from three precise surveys in Finland (Ekman 1996, Mäkinen & Saaranen 1998) (Fig. 13). On the basis of gravimetric surveys of the Geodetic Institute, the Fennoskandian land uplift is associated with some mass flow in the deep mantle layers of the Earth. When, for example, the land uplift in the Vaasa area in relation to the centre of the Earth is approximately 10 mm/y, gravity has diminished in 26 years (0.24×10×26 microgals), or about 0.06 milligals. (The Finnish Geodetic Institute, http://www.fgi.fi/yleis/historia_eng.html)

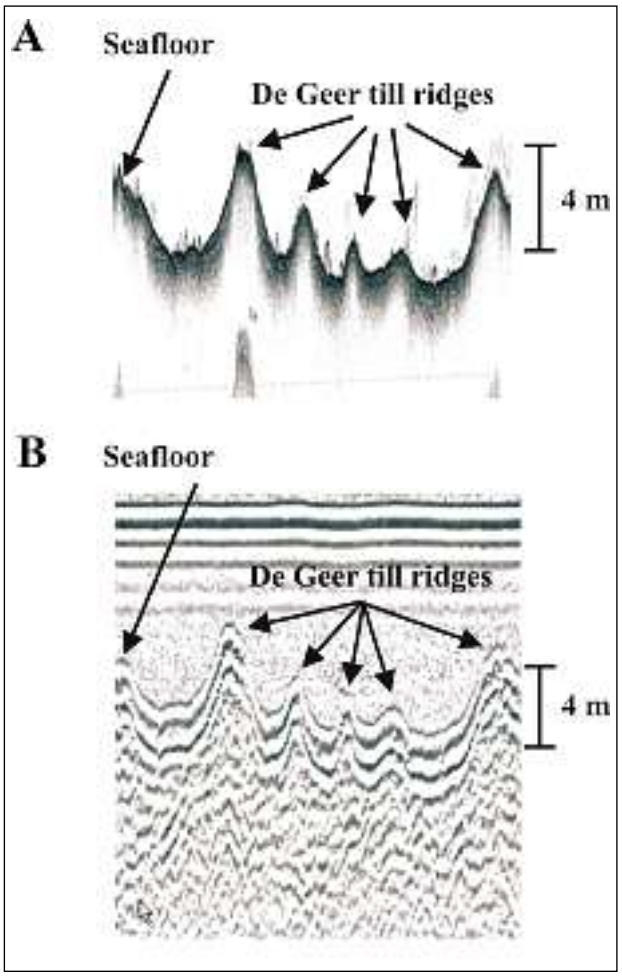


Fig. 15. Echosounding (A) and acoustic reflection profiles (B) of De Geer moraines (Compilation: A. Reijonen).

Marine geology of the Kvarken Archipelago

The narrowest part of the Gulf of Bothnia, Kvarken area, forms a submarine sill (25 m) that separates the Bothnian Sea in the south from the Bothnian Bay in the north. The Kvarken Archipelago also includes areas outside the sill where greater depths occur sparsely (~83 m at maximum). Due to the relatively rapid land uplift, the bathymetry of the Kvarken Archipelago has changed dramatically since the last deglaciation. The majority of the Kvarken Archipelago is very shallow (0–25 m) and shoaly. The fairways are shallow, boulder-rich, and mostly less than 10 meters deep. During, and just after the last deglaciation (around 10,000 years ago), the archipelago was submerged more than 200 m.

Seafloor geologic information was obtained using acoustic-seismic investigation methods and sediment sampling. Acoustic-seismic methods used include echo sounding (MeriData MD 28 kHz transmitters); single-channel seismic reflection (Electro Magnetic implosion type sound source, ELMA, 400–700 Hz, depth resolution of ± 2 m) and side scan sonar (Klein SA 350, 100 kHz) surveys.

The seafloor morphology is characterised by tectonic lines in Vaasa granite, and hummocky and De Geer moraines. The crystalline bedrock is similar on both sides of the Northern Kvarken and thus it is assumed that this is true also for the sea area, although no actual data are available (Winterhalter 2000). Sedimentary rocks exist on the seafloor of the Bothnian Sea and the Bothnian Bay but the rocks have not yet been found in the Kvarken Archipelago. However, Lower Cambrian sedimentary rocks occur in the coastal area (Söderfjärden), close to the city of Vaasa.

The entire Baltic Sea basin has undergone several

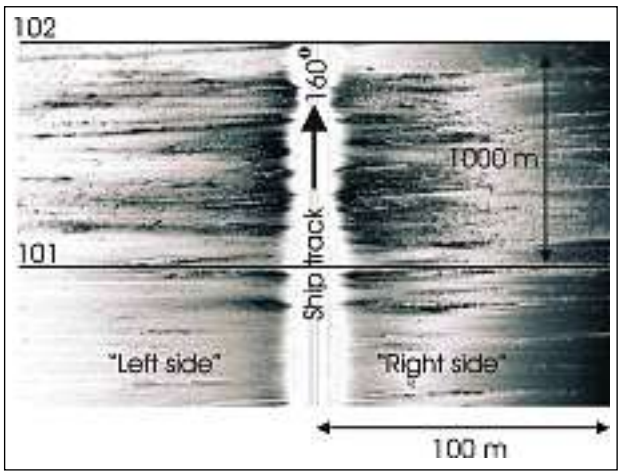


Fig. 16. Side-scan sonar profile of De Geer moraines. Side-scan sonar presents the seafloor on both sides of the vessel and produces images resembling aerial photographs.

Fig. 18. Splitting the sediment core onboard r/v Geola. White line in Figure indicates the onset of the Litorina Sea stage in lithostratigraphy. (Photo: J. Hämäläinen, 2003).

glaciations during the Late Pliocene and Pleistocene (the past ~2.7 million years). During this time, the Kvarken and Baltic Sea areas have been repeatedly subjected to glacial erosion and accumulation. However, information on possible interglacial deposits in the present marine area is still very scarce. From earlier geological stages, there are indications of land uplift in the Bothnian Bay area of 100 m above present level, and a lowering of the ocean sea level would have changed the hydrography of the whole area.

The Bothnian Bay and the entire Baltic Sea were isolated from the ocean during the Early and Mid-Weichselian sub stages, 115 000 – 50 000 years ago, when the ocean level was lower than it is today and the area was undergoing uplift after the Saalian and Early Weichselian glaciations (Lundqvist 1992, Lundqvist & Robertsson 1994, Nenonen 1995). Old preglacial river channels, tens of metres deep, have been found on the seafloor of the Bothnian Bay and the Bothnian Sea as extensions of present-day rivers (Tulkki 1977). The channels extend to the central parts of the marine area, to a depth of 80 m below present sea level, thus showing the probable ancient shoreline (Fig. 10).

The seafloor consists mainly of till. Due to strong currents and wave action, varved clay and postglacial clay sediments are often absent. Postglacial clays and gyttja-clays cover the sea floor only in basins, protected from current activity (Ignatius et al. 1980). According to the available data, the thickness of the Quaternary deposits in the Kvarken area is relatively low.

The geomorphologic feature that makes the Kvarken Archipelago unique is the occurrence of spectacular De Geer moraines (Aartolahti et al. 1995). These moraines also occur on the seafloor in the Kvarken Archipelago (e.g. Nuorteva 1988, Reijonen 2004) (Fig. 15 and 16).

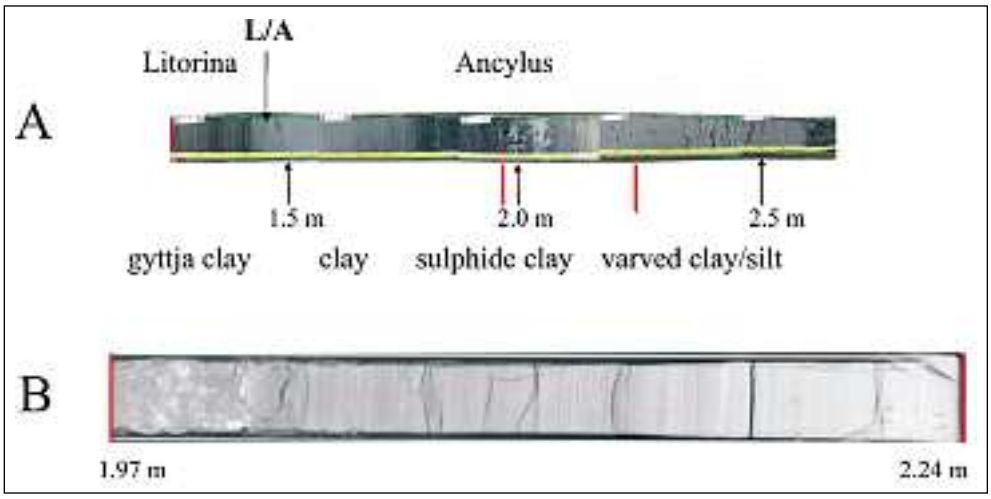


Fig. 17. Early Holocene lithostratigraphy in vibrohammer core MGTK-2003–14 of the Kvarken Archipelago (A). The onset of the Litorina Sea stage is indicated by L/A in Fig.. X-ray radiograph of sulphide clays and varved clays (1.97–2.24 metres below seafloor, red bars in Fig. A) is shown in Fig. B (Compilation: A. Kotilainen).

The glacial morphology of the seafloor has not undergone coastal deformation, as is the case with land areas. In the marine area, it is possible to study the nature of glacial features more or less in the state they were formed (Winterhalter 1972).

Despite the sparse occurrence of glacial and post-glacial clays in the Kvarken Archipelago, the Ancylus Lake and the Litorina Sea stages of the Baltic Sea are relatively well represented in submarine sediments of sheltered basins as shown in Figures 17 and 18. However, the latest history of the Baltic Sea is rarely recorded in submarine sediments of the Kvarken Archipelago.

Acknowledgements

We are very grateful to funding organizations and steering group members of Geonat – project of opportunity to write this paper: European Union (Interreg IIIA Kvarken Mitt-Skandia), Regional Council of Ostrobothnia / Saini Heikkuri-Alborzi, Vaasa town / Harri Nieminen, municipality of Korsholm / Pertti Malinen, Metsähallitus Western Finland Nature Heritage Services / Kari Hallantie, K.H.Renlunds foundation / Professor Carl Ehlers and Geological Survey of Sweden / Dr Naz Ahmed Shaikh and Anders Elhammer. Especially we thank you Susanna Ollqvist and Leena Rinkineva-Kantola of intensive work and encouraging support to geology in the Kvarken Archipelago World Heritage preparatory project and to Professor Matti Saarnisto of accepting the manuscript. We are also grateful to Harri Kutvonen from GTK who prepared many of the illustrations. The authors will thank you Hanna Virkki, Miikka Paalijärvi and Tapio Väänänen of GIS-services. Also we thank you the captain and staff of GEOLA – vessel. Finally we thank you our colleagues in GTK Jyrki Hämäläinen, Jukka Ojalainen, Sakari Kielosto, Niko Putkinen, Anu Reijonen, Seppo Töllikkö and Janne Tranberg, who made great part of field observations.

REFERENCES

Aartolahti, T. 1988. Maanpinnan muodot, Landytans former. In: Osala, T. (ed.), Vaasan saaristo, Vasa skärgård. Vaasa: O&G Förlaget/O&G Bolaget, 26–36.

Aartolahti, T. 1995. Glacial geomorphology in Finland, In: Ehlers, J., Kozarski, S. & Gibbard, P. (eds.), Glacial Deposits in North – East Europe. Rotterdam: A.A.Balkema Publishers, 37–50.

Bargel, T., Huttunen, T., Johansson, P., Jokinen, S., Lagerbäck, R., Mäkinen, K., Nenonen, K., Olsen, L., Svedlund, J.-O., Väänänen, T. & Wahlroos, J.-E. 1999. Maps of Quaternary geology in central Fennoscandia, sheet 2: glacial geomorphology and palaeohydrography. Scale 1:1 000 000. Espoo : Trondheim : Uppsala: Geological Survey of Finland : Geological Survey of Norway : Geological Survey of Sweden.

Benn, D.I. & Evans, D.J.A. 1998. Glaciers and glaciation. London: Edward Arnold, 734 p.

Breilin, O., Kotilainen, A., Nenonen, K., Virransalo, P., Ojalainen, J. & Stén, C.-G. 2004. Geology of the Kvarken Archipelago. Geological Survey of Finland, unpublished report, 47 p.

De Geer, G. 1889. Änmoräner i trakten mellan Spånga och Sunbyberg. Geologiska Föreningens i Stockholm Förhandlingar 11, 385–397.

Edlund, G.W. (ed.) 1893. Finland in the 19th century as presented by Finnish writers and artists in words and pictures, Helsinki 1893. [in Finnish]

Ekman, M. 1996. A consistent map of the postglacial uplift of Fennoscandia. Terra Nova 8, 158–165.

Eriksson, L. & Henkel, H. 1994. Geophysics. In: Fredén, C. (ed.), Geology – National Atlas of Sweden, 76–101.

Geonat 2003. Unpublished fieldwork results of the Geonat-project. Geological Survey of Finland and Department of Geology, University of Turku.

Geonat 2004. Unpublished fieldwork results of the Geonat-project. Geological Survey of Finland and Department of Geology, University of Turku.

Hoppe, G. 1957. Problems of glacial morphology and the Ice Age. Geografiska Annaler 39, 1–18.

Hätterstrand, C. 1997. Ribbed moraines in Sweden-distribution pattern and paleoglacial implications. Sedimentary Geology 11, 41–56.

Ignatius, H., Kukkonen, E. & Winterhalter, B. 1980. Pohjanlahden kvartäärikerrostumat. Geologinen tutkimuslaitos, Tutkimusraportti 45. [in Finnish]

Johansson, P. & Nenonen, J. 1991. Till stratigraphical studies in the Pulju area in northern Finland. In: Autio, S. (ed.), Geological Survey of Finland, Current Research 1989–1990. Geological Survey of Finland, Special Paper 12, 131–134.

Kakkuri, J. 1991. Planeetta Maa. Ursan julkaisuja 42. Helsinki: Tähtitieteellinen yhdistys Ursa, 184 p.

Kakkuri, J. & Virkki, H. 2004. Maa Nousee. In: Koivisto, M. (ed.), Jääkaudet. Helsinki: WSOY, 168–178.

Laaksonen, K. 1994. De Geer-moreenien syntymalleista. Terra 106, 25–34.

Linden, M., Möller, P., Adrielsson, L. & Kljaer, K. 2004. Subglasial formation of De Geer moraines, northern Sweden. Geologiska Föreniges i Stockholm Förhandlingar 126, 123 p.

Lundqvist, J. 1989. Rogen (ribbed) moraine-identification and possible origin. Sedimentary Geology 62, 281–292.

Lundqvist, J. 1992. Glacial stratigraphy in Sweden. Geological Survey of Finland, Special Paper 5, 43–59.

Lundqvist, J. & Robertsson, A-M. 1994. Glacials and Interglacials. In: Fredén, C. (ed.), Geology, National Atlas of Sweden, 120–124.

Lundqvist, J. 2000. Paleoseismicity and De Geer Moraines. Quaternary International 68, 175–186.

Mawdsley, J.B. 1936. The wash-board moraines of the Opawica-Chibougamau area Quebec. Transaction of the Royal Society of Canada 30, Ser. III, 9–12.

Mäkinen, J. & Saaranen, V. 1998. Determination of post-glacial land uplift from three precise leveling in Finland. Journal of Geodesy 72, 516–529.

Alalammi, P. (ed.) 1990. Atlas of Finland. Folio 123–129: Geology. Helsinki: National Board of Survey, Geographical Society of Finland, 58 p.

Nenonen, K. 1995. Pleistocene stratigraphy and reference sections in southern and western Finland. Geological Survey of Finland, Miscellaneous Publications, Ph.D. Thesis, 205 p.

Nuorteva, J. 1988. Akustisilla luotausmenetelmillä saatu kuva

merenpohjan kvartäärikerrostumista. Geologian tutkimuskeskus, Tutkimusraportti 82, 32 p. [in Finnish]

Ollqvist S. & Rinkineva-Kantola, L. (eds.) 2004. Nomination of the Kvarken Archipelago for the inclusion in the World Heritage list. Metsähallitus Western Finland Natural Heritage Services, West Finland Regional Environment Centre, Regional Council of Ostrobothnia, Vaasa, 85 p.

Osala, T. 1988. Vaasan Saaristo – Vasa Skärgård. Vaasa: O&G Förlaget, 424 p.

Palomäki, M. 1988. Maankohoaminen, Landhöjningen. In: Osala, T. (ed.), Vaasan Saaristo, Vasa Skärgård, Vaasa: O&G Förlaget/ O&G Bolaget, 37–65.

Reijonen, A. 2004. Vedenalaisten moreenimuodostumien monimuotoisuus Merenkurkussa. University of Helsinki, Department of Geology, Unpublished M.Sc. Thesis, 102 p. [in Finnish]

Saarnisto, M. 1981. Holocene emergence history and stratigraphy in the area north of the Gulf of Bothnia. Annales Academiae Scientiarum Fennicae. Series A. III. Geologica – Geographica 130, 42 p.

Sarala, P. 2003. Ribbed-moreenit – jäätikön liikesuunnan poikittaiset indikaattorit (Summary: Ribbed moraines – transverse indicators of the ice flow direction). Geologi 55, 250–253. [in Finnish]

Sarala, P. 2004. Deglaciation pattern and glacial dynamics of the southern Finnish Lapland. In: Mansfeld, J. (ed.) The 26th Nordic Geological Winter Meeting, January 6th – 9th 2004, Uppsala, Sweden, abstract volume. GFF 126, p. 124.

Seifert, T., Tauber, F. & Kayser, B. 2001. A high-resolution spherical grid topography of the Baltic Sea revised edition. Proceedings of the Baltic Sea Science Conference 2001, Stockholm, 25–29.

Svendsen, J.I., Alexanderson, H., Astakhov, V.I., Demidov, I., Dowdeswell, J.A., Funder, S., Gataullin, V., Henriksen, M., Hjort, C., Houmark-Nielsen, M., Hubberten, H.W., Ingólfsson, Ó., Jakobsson, M., Kjær, K.H., Larsen, E., Lokrantz, H., Lunkka, J.P., Lyså, A., Mangerud, J., Matiouchkov, A., Murray, A., Möller, P., Niessen, F., Nikolskaya, O., Polyak, L., Saarnisto, M., Siegert, C., Siegert, M.J., Spielhagen, R.F. & Stein, R. 2004. Late Quaternary ice sheet history of northern Eurasia. Quaternary Science Reviews 23, 1229–1271.

Taipale, K. & Saarnisto, M. 1990. Tulivuorista jääkausiin: Suomen maankamaran kehitys. Porvoo: WSOY, 416 p. [in Finnish]

Tulkki, P. 1977. The bottom of the Bothnian Bay. Geomorphology and sediments. Merentutkimulaitoksen julkaisuja 241, 5–89.

Winterhalter, B. 1972. On the geology of the Bothnian Sea, an epeiric sea that has undergone Pleistocene glaciation. Geological Survey of Finland, Bulletin 258, 66 p.

Winterhalter, B. 2000. Sedimentary rocks underlying the Gulf of Bothnia. In: Lundqvist, T. & Autio, S. (eds.), Description to the bedrock map of central Fennoscandia (Mid-Norden). Geological Survey of Finland, Special Paper 28, 76–79.

Quaternary studies in the northern and Arctic regions of Finland
Edited by Antti E.K. Ojala
Geological Survey of Finland, Special Paper 40, 113–121, 2005.

HISTORY AND PREHISTORY OF LAKE VETSIJÄRVI

by
Tuija Rankama[1)] & Jarmo Kankaanpää[2)]

[1)]*Department of Archaeology, University of Helsinki Institute of Cultural Studies, P.O. Box 59, FIN-00014, University of Helsinki, FINLAND*
[2)] *Isonpellonkuja 6, FIN-02880, Veikkola, FINLAND*
E-mail: tuija.rankama@helsinki.fi, jarmo.kankaanpaa@kolumbus.fi

Rankama, Tuija & Kankaanpää, Jarmo, 2005. History and Prehistory of Lake Vetsijärvi. *Geological Survey of Finland, Special Paper 40,* 113–121, 9 figures.

An archaeological survey carried out in 2002 around Lake Vetsijärvi, Utsjoki parish, turned up twelve Stone-Age sites, nine of which were located more than six metres higher than the present lake. The placement of the sites and certain other observations suggest that the level of Lake Vetsijärvi may have been higher during the Stone Age than it is today. One of the sites produced finds connecting it with the earliest occupation of the northern Norwegian coast. A test excavation at this site in 2004 recovered evidence of a core-and-blade industry based on foreign chert-like raw material. This type of industry is previously known from the Varangerfiord in Norway but is unique in Finland. Finds include blades, cores, and a tanged arrowpoint with parallels in northern Russia.

Key words (GeoRef Thesaurus, AGI): archaeological sites, lakes, paleolimnology, lake-level changes, artifacts, Stone Age, Lake Vetsijärvi, Utsjoki, Finland

Introduction

Lake Vetsijärvi is the largest lake in the Kaldoaivi region, the eastern wilderness of Utsjoki, northern Finnish Lapland (Fig. 1). In 2002, the authors carried out an archaeological survey at the lake (Rankama & Kankaanpää 2003). The results of the survey gave rise to a few intriguing ideas about the history of the lake. They also led to an excavation at one of the sites, producing a finds assemblage not previously encountered in the inland regions of northern Scandinavia.

The Survey

The purpose of the survey was to test a model, previously presented by Rankama (1996), according to which the fish resources of the inland lakes in Utsjoki parish played an important role in the prehistoric economy of the region and could have been actively exploited as early as the Mesolithic Period. Lake Vetsijärvi was chosen for the survey because it is rich in fish (Niemelä & Vilhunen 1987) and would thus have been among the most likely locations for seasonal occupation. It is also one of the largest lakes in Utsjoki, with a surface area of ca. 8.2 km^2. The lake lies in a shallow depression between fell chains and is surrounded by an extensive area of bogs. It is fed by several small rivers that drain the plateau particularly north and east of the lake. Lake Vetsijärvi empties into the Teno River via the Vetsijoki River, which begins at the lake's northwestern corner and – after meandering for a while – runs nearly due north to meet the Teno at Vetsikko.

Since the water level of lakes in northern Lapland is

Fig. 1. Location of the research area.

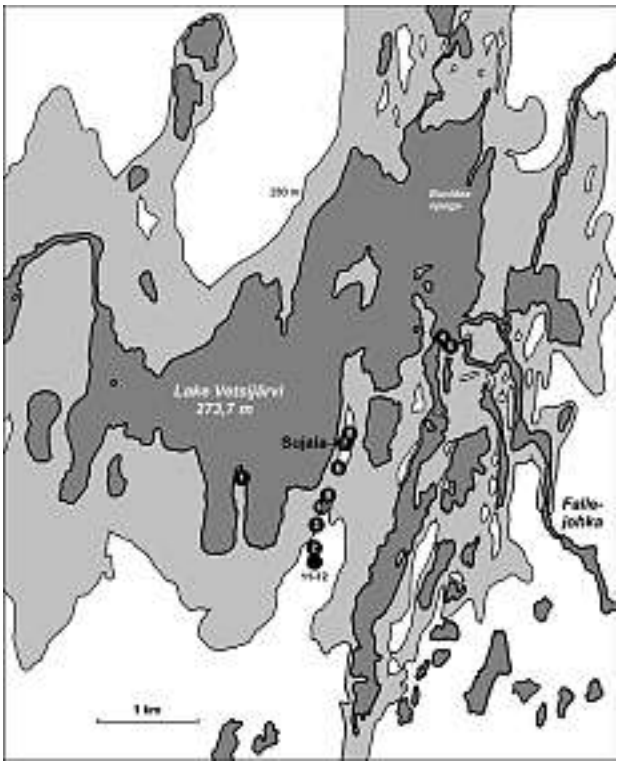


Fig. 2. Sites located during the survey of 2002. The Sujala site is nr. 7.

considered to have risen after the end of the Postglacial Climatic Optimum (e.g. Hyvärinen & Alhonen 1994, Eronen et al. 1999, Eronen et al. 2002), it was hoped that a survey of the shoreline would turn up inundated Stone Age sites. The fact that the shores of Lake Vetsijärvi are largely boggy presented the additional possibility that shoreline sites might be partly covered by peat, which would offer an anaerobic environment with a chance of organic preservation – something otherwise extremely rare in Finnish archaeology.

Several kilometres of the Lake Vetsijärvi shoreline were surveyed by kayak in the summer of 2002 for the purpose of locating shoreline sites, but the rewards were few: only three sites were discovered near the shore, and only one of these was actually in the water. However, nine further sites were discovered along a dirt track running the length of a peninsula protruding into the lake from the south (Fig. 2). All of these latter sites were located above the 280-metre mark, or a good six metres above the present level of the lake, which in early July 2004 lay at an elevation of 273.72 m above sea level. Horizontally, the sites lay a minimum of 80 metres from the present lakeshore, in many cases much further. All of the sites – with one exception – were marked by the presence of quartz flakes, a characteristic feature of Finnish Stone Age and Early Metal Age sites. The one exception, the site named *Utsjoki 226 Vetsijärvi 7 Sujala*, produced material that pointed in another direction but also suggested an exceptional age. We will return to this site later in the paper.

The Water Level Question

There appears to be no practical reason why the campsites should have been located far from shore. Level and dry shoreline locations with sandy beaches – the type of setting usually preferred by Stone Age hunter-gatherers – are available nearby, at the base of the peninsula and at its tip, but upon inspection these sites revealed no signs of Stone Age occupation. Spring flooding does not appear to affect the lake to any great extent, at least judging from the fact that a cabin in the southeastern part of the lake stands on top of a geodetic benchmark with an elevation of 275.7 m, or a scant two metres above the early July level.

Locating the campsites on high ground rather than near the shore makes little sense from a use perspective. If the camps related to fishing, the natural place would have been near the shore to minimize carrying distances. If they related to reindeer hunting, the whole location is wrong. Higher, better, and more easily hidden lookout posts could have been established on the hills to the south. A campsite located on or near the base of the peninsula would also preclude the possibility of using the peninsula itself as a natural "trap" where reindeer would have been lured by the relatively insect-free windy location or where they could have been actively driven by beaters.

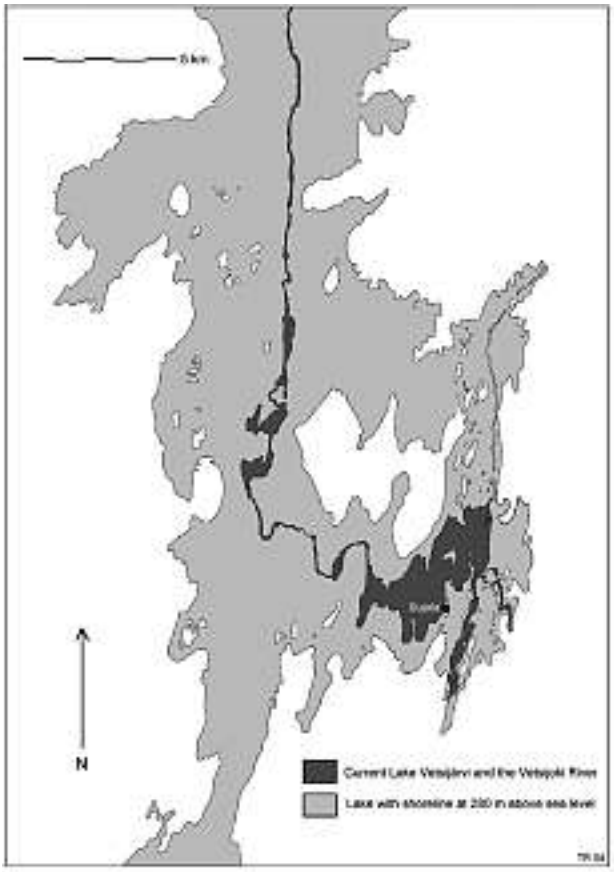


Fig. 3. Potential size of Lake Vetsijärvi with water level at 280 metres.

The most logical remaining explanation for the clustering of nine sites above the 280-metre contour would seem to be that the sites were in fact located close to the shoreline *in their time*, in other words the level of the lake during at least part of the Stone Age was actually higher relative to the surrounding area than it is today. Plotting the shoreline on the 280 -metre contour (Fig. 3) gives a much larger lake and locates many of the sites on an island separated from the mainland by a narrow and shallow strait.

A higher lake level during the Stone Age would, however, be contrary to the apparent general trend in the region. For this reason, future research at Lake Vetsijärvi should include geological studies on the lake's history. Phenomena observed during previous research that might figure in such studies include the following:

Thick peat layers in elevated dry areas

A layer of peat more than 50 cm thick was observed on top of a very narrow and steep-sided moraine ridge that forms a peninsula in the northeastern corner of the lake (Fig. 4). The ridge rises an estimated 5–6 metres above the lake. Due to the coarse soil and the sharp contours of the ridge, the ground appears to be well drained. Since the formation of peat usually requires a large amount of moisture, the thick peat layer on top of the ridge might suggest that the water level has been higher.

Possible beach formations

A formation resembling an eroded shoreline terrace runs along the western side of the peninsula where most of the sites are located; the eastern side has not been surveyed. This formation has not been systematically mapped, but it can be observed in the surface profile plotted from the vicinity of the northern find cluster of the Sujala site mentioned above down to the

Fig. 4. Thick peat on top of the *Buolžanjarga* ridge. (Photo: T. Rankama, 2002)

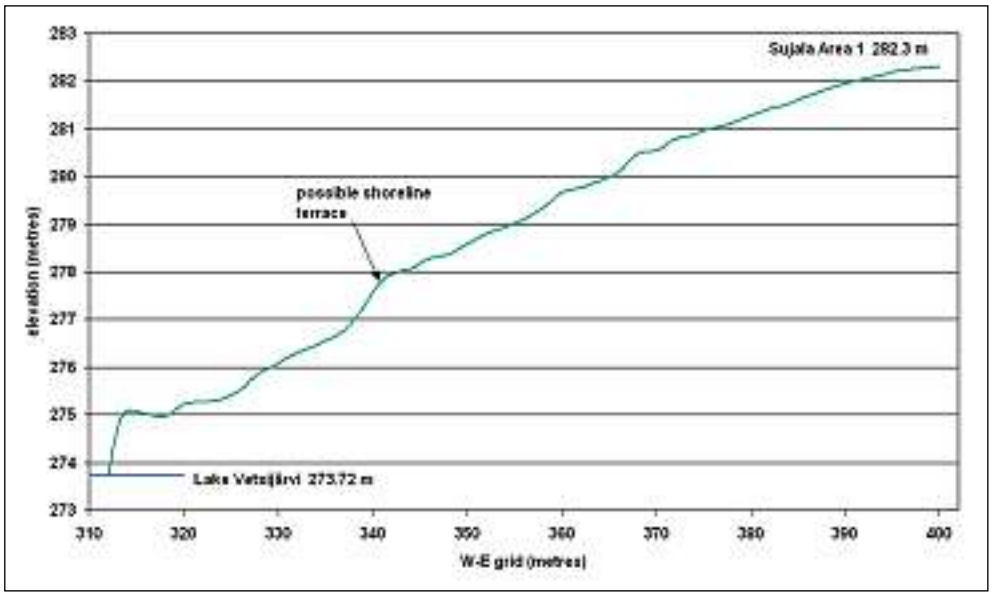


Fig. 5. Surface profile from the Sujala site.

lakeshore (Fig. 5) and also as a vegetation border on an infrared aerial photograph of the area. The elevation of the feature in the surveyed profile is ca. 278.0 metres for the edge of the "terrace" and 276.8 metres for the foot of the underlying slope. The latter elevation is more than a metre higher than the benchmark under the cabin. Two smaller "terrace-like features" may be observed in the surface profile at elevations of 279.7 and 280.5 metres, but these have not been investigated as of yet and may be just natural bumps in the terrain. One further possible indication of an ancient beach line was noted in connection with the test excavation of the Sujala site in 2004: a concentration of rocks found in the lowest test square of the northern cluster at an elevation of ca. 282.0 metres. The other – higher – test squares in this area contained mixed sand and gravel with interspersed larger stones, but this square contained almost pure rocks reminiscent of a water-washed beach line.

Tilting basin

The topography of the Lake Vetsijärvi area itself is rather flat. Though the lake is surrounded by fells, the present 280-metre elevation curve outlines a vast area that appears to continue both north and south. The present outlet of Lake Vetsijärvi is to the north through the Vetsijoki River, but isostatic rebound has tilted the basin northward (Eronen 1979) and it is consequently conceivable that the original postglacial outlet may have been towards the southwest and the Utsjoki River. Geological and topographical studies are needed to establish whether it would have been possible for the level of the lake to have been higher (relative to the present topography) when the surface was tilted more towards the south.

The Sujala Site

The most intriguing of the sites discovered during the 2002 survey was number 7, dubbed Sujala (officially, *Utsjoki 226 Vetsijärvi 7 Sujala*) after the cabin owner in whose "backyard" it was found. As with most of the other sites, the evidence of prehistoric human activity consisted of a number of lithic flakes and flake fragments discovered on the track leading to the lake from the south. What made the Sujala site interesting was the fact that both the raw material and the shape of the flakes differed from those of the other Vetsijärvi sites and from those usually found at Stone Age sites in Finland.

Instead of quartz, which dominates the Finnish Stone Age, the flakes were made from a variety of raw materials more common on the North Norwegian coast, such as quartzite, flint, and chert. The shape of the flake fragments suggested a specialized knapping technology geared towards the production of long, slender, parallel-sided flakes called blades. This technology is alien to Finnish Stone Age contexts and its products are usually considered to have been imported. In northern Norway, the technology is considered Mesolithic, i.e., part of the earliest Stone Age of the area (Olsen 1994, Woodman 1993, 1999).

Because of these unusual finds, a test excavation was

Fig. 6. Blade cores from Sujala. (Photo: J. Kankaanpää, 2004)

carried out at Sujala in 2004 (Kankaanpää & Rankama 2004). A careful surface survey at the beginning of the excavation revealed two main clusters of finds on the track some 200 metres apart. Eight 1 x 1 metre test squares were excavated around each of the clusters. A total of 379 artefacts were recovered, both from the track surface (50.7 %) and from the ten test squares closest to the track (49.3 %). The clusters of finds were small in diameter and close to the soil surface, suggesting short-term settlement, probably small single-occupation campsites (see Rankama & Kankaanpää in press for more details of the excavation).

The excavation finds bore out the unusual character of both raw material use and technology at Sujala. Only eight per cent of the finds were quartz, while the dominating raw material (87 %) was a light brown chert-like rock. This rock, which used to be erroneously called "dolomite" by Norwegian archaeologists, was commonly used during the Stone Age along the coast of the Varangerfjord (e.g. Simonsen 1961: *passim*). The current Norwegian name for it is "tuffaceous chert" (e.g. Hood 1992).

Microscopic surface examination reveals a very fine-grained matrix with even grain size. This suggests that the material does not originate in the Fennoscandian Shield; the northern Norwegian Caledonides are a more probable source area (J. Välimaa, T. Manninen & R. Kesola personal communication 2005). Pending a proper mineralogical analysis, we will call the rock simply "chert". It is fairly easily worked and forms very sharp edges. Its colour is dark grey or greenish when fresh, but it weathers to a tan or light grey colour and becomes brittle in the process. This raw material is very rare at Stone Age sites in Inari and Utsjoki (e.g., Kankaanpää & Rankama in press). Because of the commonness of the raw material in Stone Age contexts along the Varangerfjord, the few discoveries from Finnish Lapland have been considered imports from the coast (e.g. Havas 1999).

Chert had been used at Sujala to produce blades from carefully shaped blade cores. Two exhausted cores were discovered, one conical in shape (Fig. 6:1), the other a one-sided core with two opposing platforms and acute platform angles (Fig. 6:2). The blades produced from these cores during their final stage were not very large; however, the size of the recovered blade fragments (e.g. Fig. 7) indicates that the cores were originally much larger, or at least that a large number of the blades were produced from larger cores.

The total number of recovered blades and blade fragments, including implements, was 118. In addition, more than 30 core tablets, i.e., platform rejuvenation flakes, were recovered (Fig. 8). These are flakes produced during blade production, when the striking platform of the core needs to be renewed because of damage to the platform edge or because of the increasing curvature of the core face due to continued blade production. Since these flakes can hardly be used as implements of any kind, they are the most unquestionable evidence that blades were, indeed, produced at the Sujala site and not just brought there from somewhere else.

About 50 % of the chert artefacts from Sujala

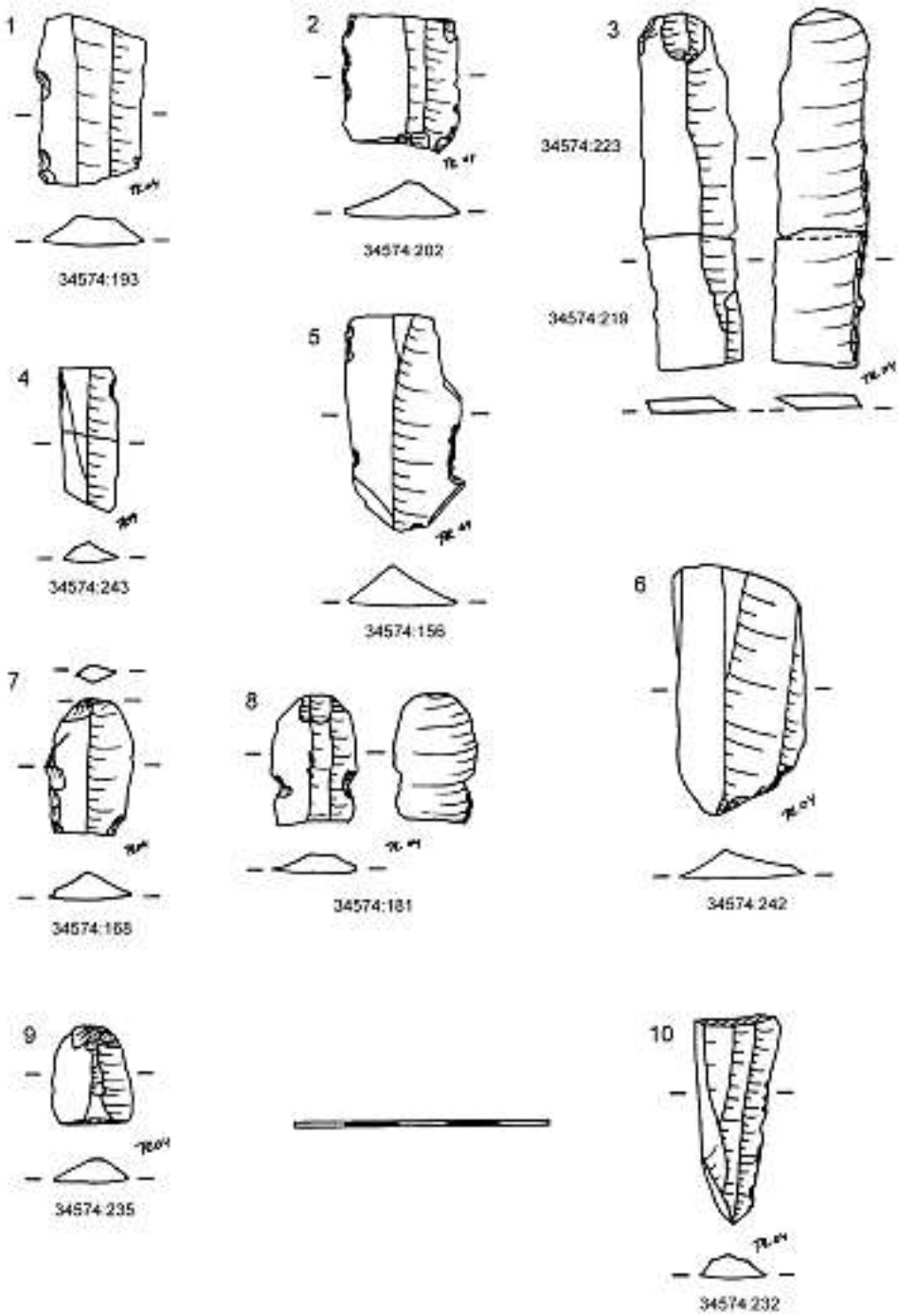


Fig. 7. Blade fragments from Sujala.

relate indisputably to blade technology. The rest of the artefacts consist mostly of small debris classified as flake fragments because they are too small to be reliably identified as deriving from blade production. The majority of them, however, probably do.

The Sujala assemblage includes one tanged point made of chert (Fig. 9). It is 42.1 mm long and made from a large blade so that the central dorsal ridge of the blade runs along the centre of the point's long axis. The tang is bifacially flaked and is diamond-shaped in cross section; the tip has invasive retouch on its ventral side, making its cross section also diamond-shaped.

The combination of raw material and technology at Sujala is alien to Finnish archaeology. Since the site lies in the inland region and has never been part of the coastal sphere, it cannot be dated through shore displacement chronology. The radiocarbon samples

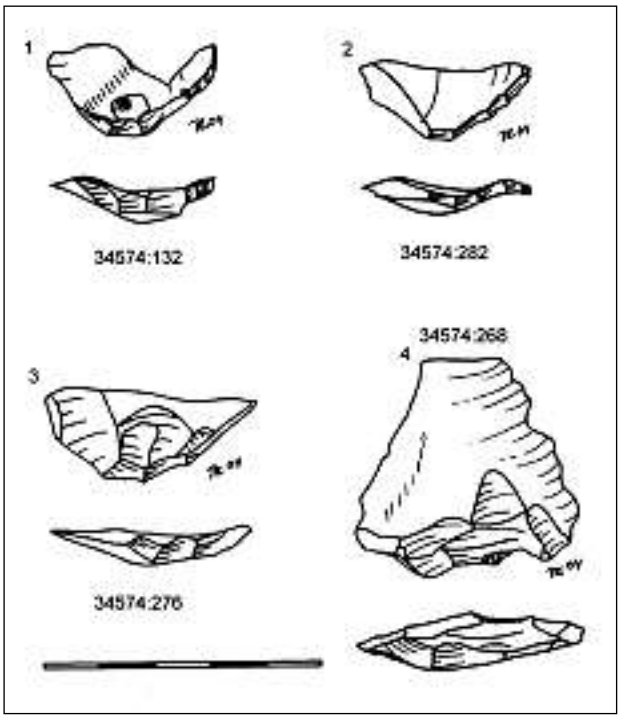


Fig. 8. Core tablets from Sujala.

collected so far are too equivocal for reliable analysis. Therefore, the only viable option is typological dating.

The nearest area where a comparable large blade technology exists is the North Norwegian coast, where it is dated to the Preboreal, representing the earliest Stone Age of the region. If the Sujala occupation is associated with the coastal sphere, it is the first discovered inland settlement of Phase I of the North Norwegian Mesolithic, also referred to as the Komsa Culture (see Olsen 1994, Woodman 1993, 1999). It thus changes our view of the adaptation of the early Komsa population, which has traditionally been considered completely maritime. At the same time, it changes the chronology of the settlement of northern Finnish Lapland: the earliest dates so far derive from the Saamen Museo site in Inari, where two pieces of birch bark pitch have been dated to 8835±90 BP and 8760±75 BP (Carpelan 2003). Phase I of the North Norwegian Mesolithic is dated to 10 000–9000 BP (Olsen 1994).

The tanged point from Sujala, however, fits poorly in the Komsa sphere, where the tanged points are usually smaller, more irregular, and never have invasive retouch on the ventral side. Instead, the point is very well at home among the Early Mesolithic so-called Post-Swiderian complexes of Northwest Russia (e.g. Sorokin 1984, Volokitin in press, Zhilin 1996), and so is the sophisticated blade technology as a whole. These complexes are also Preboreal, but they continue into the Boreal period. If this is the origin of the Sujala population, the site is very far from the core areas of these complexes (e.g. Burov 1999) – the nearest site with Post-Swiderian type material is Ristola in Lahti, southern Finland (Takala 2005), itself an outlier. The routes taken by the Post-Swiderian people to reach Utsjoki, in that case, are totally obscure. It also remains a mystery how, representing as they did an inland adaptation, they came to find the raw material so typical of the coastal assemblages.

Conclusions

The survey and excavation demonstrated that the Lake Vetsijärvi region had indeed been visited by

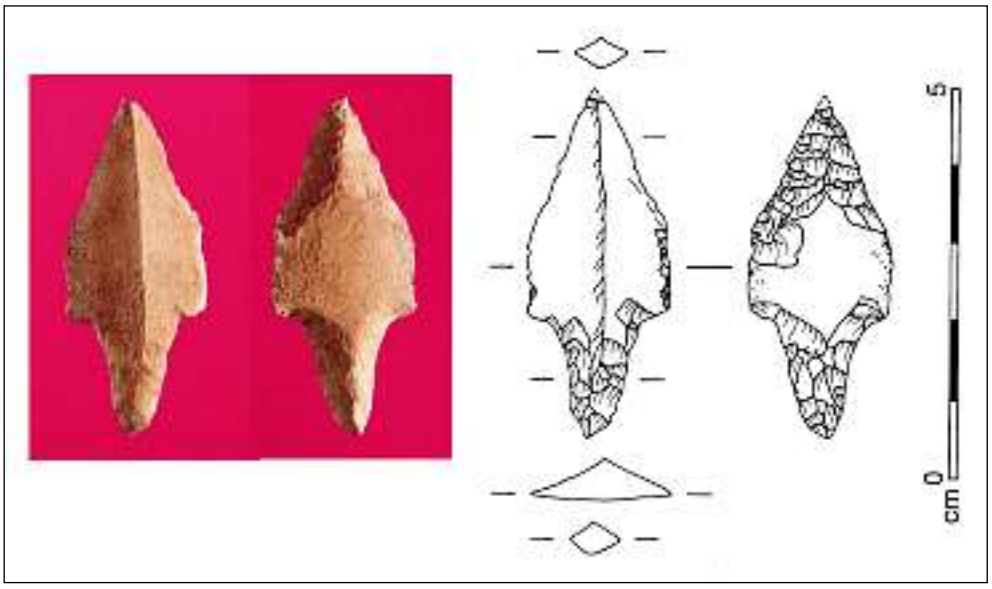


Fig. 9. Tanged point from Sujala. (Photo: J. Kankaanpää, 2004)

people during the Stone Age, although the details of these visits – seasonality, utilized resources, group size etc. – still require further research. Very little evidence of occupation was found on or near the present shoreline in spite of the fact that most of the shoreline of the main basin was carefully inspected. Unless the water level of the lake has risen several metres since the end of the Stone Age and thus covered even the latest sites, one would have expected to find evidence of human presence in places where the mineral soil lay bare on the shoreline, particularly on sandy beaches. A rise in the lake level, however, would not account for the fact that a notable cluster of sites was discovered some 6–7 metres higher than the present shoreline. In southern Finland, such a cluster would automatically be connected to shore displacement. At Lake Vetsijärvi, a lowering of the water level would go against the current view of lake history in the region. The rise of lake levels after the Climatic Optimum has been connected with increased precipitation, but in the case of the Vetsijärvi basin the effects of the cooling trend may have been nullified by isostatic tilting, which could have opened a new outlet, as happened with the Päijänne and Saimaa basins in Southern Finland. A rise of the lake level could in itself have also caused the opening and rapid erosion of a new exit channel, leading to a quick drop in the water level (M. Eronen & M. Saarnisto, personal communication 2005). Geological studies of Lake Vetsijärvi are necessary for solving this question.

The Sujala site represents the earliest human occupation recorded so far in the Lake Vetsijärvi region and possibly in all of Finnish Lapland. The origin of the occupation remains equivocal at the moment, but whichever of the possible regions – coastal North Norway or Northwest Russia – it came from, it has the potential for profoundly changing archaeologists' views concerning the colonization processes and settlement systems of northern Lapland. A coastal origin would indicate that the adaptation of the Early Komsa population was not so purely maritime as has been thought so far. On the other hand, an origin in Northwest Russia would force a rethinking of the way the north was settled as a whole: it would no longer be possible to assume that all the settlers travelled along the Norwegian coast from the south and ultimately from Northwest Europe. On a smaller scale, the site also changes the chronology of the settling of Finnish Lapland, pushing it back in time potentially several hundred years.

The history and prehistory of Lake Vetsijärvi and its surroundings obviously merit further intensive study.

Acknowledgments

The test excavation of the Sujala site would not have been possible without the help of our volunteer Scandinavian field crew: Ericka Engelstad, Helena and Kjel Knutsson, Mikael Manninen, Lise Sand, and Taarna Valtonen. We wish to thank Mr. Eero Sujala of Puksala, Utsjoki for transporting us to and from Vetsijärvi both in 2002 and 2004, and for the use of his cabin. Thanks are also due to Matti Seppälä, Jukka Välimaa, Tuomo Manninen and Reino Kesola for their help during the writing of this paper and to Matti Eronen and Matti Saarnisto for their reviews and useful comments. Tuija Rankama's work in connection with the excavation and the analysis and writing was financed by the Finnish Cultural Foundation.

REFERENCES

Burov, G.M. 1999. "Postswiderian" of the European north-east. In: Tanged Points Cultures in Europe. Lublin: Maria Curie-Skłodowska University Press, 281–291.

Carpelan, C. 2003. Arkeologiset vaiheet. In: Lehtola, V.-P. (ed.), Inari – Aanaar. Inarin historia jääkaudesta nykypäivään. Oulu: Inarin kunta, 30–95. [in Finnish]

Eronen, M. 1979. The retreat of pine forest in Finnish Lapland since the Holocene climatic optimum: a general discussion with radiocarbon evidence from subfossil pines. Fennia 157, 93–114.

Eronen, M., Hyvärinen, H. & Zetterberg, P. 1999. Holocene humidity changes in northern Finnish Lapland inferred from lake sediments and submerged Scots pines dated by tree-rings. The Holocene 9, 569–580.

Eronen, M., Zetterberg, P., Briffa, K.R., Lindholm, M., Meriläinen, J. & Timonen, M. 2002. The supra-long Scots pine tree-ring record for Finnish Lapland: Part 1, chronology construction and initial inferences. The Holocene 12, 673–680.

Havas, H. 1999. Innland uten landegrenser. Bosetningsmodeller i det nordligaste Finland og Norge i perioden 9000–6000 BP. Hovedfagsoppgave, Institutt for arkeologi, Universitetet i Tromsø. 147 p.

Hood, B.C. 1992. Prehistoric Foragers of the North Atlantic: Perspectives on Lithic Procurement and Social Complexity in the North Norwegian Stone Age and the Labrador Maritime Archaic. Ph.D. dissertation, University of Massachusetts. Ann Arbor: UMI Dissertation Services, 564 p.

Hyvärinen, H. & Alhonen, P. 1994. Holocene lake-level changes in the Fennoscandian tree-line region, western Finnish Lapland: diatom and cladoceran evidence. The Holocene 4, 249–256.

Kankaanpää, J. & Rankama, T. 2004. Utsjoki 226 Vetsijärvi 7 Sujala. Varhaismesoliittisen asuinpaikan koekaivaus 27.6.–3.7.2004. Unpublished test excavation report. Copy kept at the archive of the National Board of Antiquities Department of Archaeology.

Kankaanpää, J. & Rankama, T. Pioneer settlement in northern Finnish Lapland. In: H. Knutsson (ed.), Pioneer settlements and colonization processes in the Barents region. First NordArk publication. Vuollerim, in press.

Niemelä, E. & Vilhunen, J. 1987. Utsjoen tunturivesien kalakantojen käyttö- ja hoitosuunnitelma. Helsinki: Riista- ja kalatalouden tutkimuslaitos, kalantutkimusosasto, 187 p.

Olsen, B. 1994. Bosetning og samfunn i Finnmarks forhistorie. Oslo: Universitetsforlaget, 158 p.

Rankama, T. 1996. Prehistoric Riverine Adaptations in Subarctic Finnish Lapland: The Teno River Drainage. Ph.D. dissertation, Brown University Department of Anthropology. Ann Arbor: UMI Dissertation Services, 955 p.

Rankama, T. & Kankaanpää, J. 2003. Utsjoki Vetsijärvi. Arkeologinen inventointi 18.–23.7.2002. Unpublished survey report. Copy kept at the archive of the National Board of Antiquities Department of Archaeology, 34 p.

Rankama, T. & Kankaanpää, J. Survey and excavation at Lake Vetsijärvi, Lapland. Proceedings of the 22nd Nordic Archaeology Conference, Oulu, Finland, August 2004, in press.

Simonsen, P. 1961. Varanger-funnene II. Fund og udgravinger på fjordens sydkyst. Tromsø Museums Skrifter VII/II, 524 p.

Sorokin, A.N. 1984. Mezolit Velikikh Meshtsherskikh Ozer. Sovetskaya Arkheologiya 1984/1, 46–65.

Takala, H. 2005. The Ristola Site in Lahti and the Earliest Postglacial Settlement of South Finland. Lahti City Museum, 205 p.

Volokitin, A.V. Some Peculiarities of Colonization of the European North-East in Mesolithic. In: H. Knutsson (ed.), Pioneer settlements and colonization processes in the Barents region. First NordArk publication. Vuollerim, in press.

Woodman, P.C. 1993. The Komsa Culture, a Re-examination of its Position in the Stone Age of Finnmark. Acta Archaeologica 63, 57–76.

Woodman, P.C. 1999. The Early Postglacial Settlement of Arctic Europe. In: Cziesla, E., Kersting, Th. & Pratsch, St. (eds.), Den Bogen spannen Festschrift für Bernhard Gramsch zum 65. Geburtstag. Beiträge zur Ur- und Frühgeschichte Mitteleuropas 20. Weissbach: Beier & Beran, 297–312.

Zhilin, M.G. 1996. The Western Part of Russia in the Late Palaeolithic – Early Mesolithic. In: Larsson, L. (ed.), The Earliest Settlement of Scandinavia and its relationship with neighbouring areas. Acta Archaeologica Lundensia Series in 8°, No. 24, 273–284.

Quaternary studies in the northern and Arctic regions of Finland
Edited by Antti E.K. Ojala
Geological Survey of Finland, Special Paper 40, 123–130, 2005.

CHEMICAL CONSTRAINTS ON THE INFLUENCE OF SUBSURFACE AND SURFACE MELTING ON LAKE BASEN, WESTERN DRONNING MAUD LAND, ANTARCTICA

by
Anne Lehtinen[1)] & Arto Luttinen[2)]

[1)]*Posiva Oy, FIN-27160, Olkiluoto, FINLAND*
[2)]*Department of Geology, P.O. Box 64, FIN-00014, University of Helsinki, FINLAND*
E-mail: anne.lehtinen@posiva.fi, arto.luttinen@helsinki.fi

Lehtinen, Anne & Luttinen, Arto, 2005. Chemical constraints on the influence of subsurface and surface melting on Lake Basen, western Dronning Maud Land, Antarctica. *Geological Survey of Finland, Special Paper 40*, 123–130, 6 figures and 1 table.

Solar radiation-induced subsurface melting has been suggested to account for frozen lakes, i.e. blue ice areas characterized by a liquid interlayer, in the coastal regions of Antarctica. Lake Basen, one of such frozen lakes and located adjacent to a major nunatak in Vestfjella, Dronning Maud Land, developed a distinctive (≤1.3 m thick) liquid layer during the austral summer of 2003–2004. Thickening of the liquid layer was coupled with a 4-fold increase in ion concentrations (Na^+, Ca^{2+}, Mg^{2+}, Cl^-, SO_4^{2-}). The lake water can be distinguished from Antarctic snow by its different ion ratios, specifically by its high Ca^{2+}/Mg^{2+}, which show a strong affinity to pond and stream water on the nunatak. Surface melting on nunatak Basen has a profound influence on the lake water chemistry, even though the proportion of ion-concentrated surface melts may be limited to a few percent. We associate the source of Ca^{2+} with the abundant carbonate precipitates on Basen. Concentrations of K^+ and NO_3^- remained notably low in the lake, presumably due to consumption by photosynthetic organisms in the drainage area.

Key words (GeoRef Thesaurus, AGI): ice sheets, thawing, hydrology, frozen lakes, hydrochemistry, ions, Lake Basen, Antarctica

Introduction

During the summer season, solar radiation-induced melting leads to the development of supraglacial lakes in the coastal areas of Greenland and Antarctica. The supraglacial lakes of Greenland are generated by surface melting and streaming and the accumulation of water in glacial depressions (Echelmeyer et al. 1991). In contrast, the supraglacial lakes of Antarctica, which are commonly referred to as frozen lakes, are associated with blue ice areas near glacial slopes and nunataks (Winther et al. 1996, Bintanja 1999, Liston et al. 1999a, 1999b). In Antarctica, the temperature of the air-ice interface and the subsurface temperature of the ice also remain below the freezing point in the summer. According to Brandt & Warren (1993), short-wave solar radiation penetrates the blue ice, which has a low albedo. The absorption of radiation produces thermal energy, which leads to the generation of a layer of liquid water within the ice. The subsurface melting process has been referred to as solid state greenhouse effect (Brandt & Warren 1993). Towards the midsummer, the subsurface liquid layer gradually becomes thicker. Emission of long-wave solar radiation from the ice surface keeps the air-water interface frozen (Brandt & Warren 1993) and the thickness of the ice cover typically exceeds 15 cm (Liston et al. 1999b). The liquid layer has been reported to attain a thickness

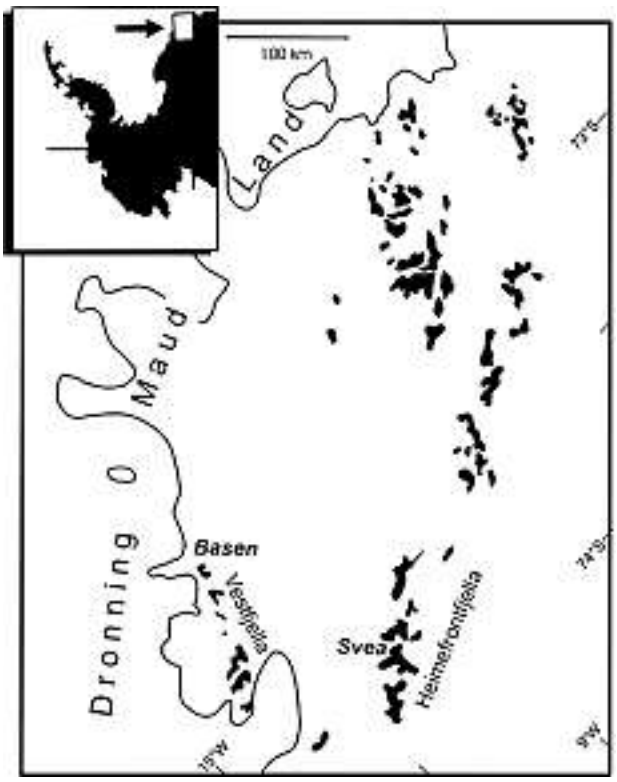


Fig. 1. Location of the study area (Basen) in north Vestfjella, western Dronning Maud Land, East Antarctica. Black areas indicate exposed bedrock. Heimefrontfjella is located on the East Antarctic plateau escarpment.

of 0.5 to 1 m in the midsummer; in winter, the lakes are completely frozen (Liston et al. 1999b).

The frozen lakes of Antarctica have so far remained fairly little studied. Many of the lakes are located adjacent to nunataks (Liston et al. 1999b). We consider that the subsurface melting model may not be directly applicable to such frozen lakes, because surface melting is enhanced on bedrock outcrops. In the subsurface melting model, frozen lakes that are located away from nunataks result from the melting of blue ice *in situ* (Bøggild et al. 1995, Liston et al. 1999a). Given the negative mass balance and the consequent exposure of ancient ice in blue ice areas (Bintanja 1999), the composition of the melt layer should correspond to that of the ascending old ice. By inference, a melt layer that is generated by subsurface melting should be chemically more akin to snow that is accumulated further inland rather than the compositionally different, near-coastal snow-cover in the vicinity of the frozen lake (Stenberg et al. 1998, Isaksson et al. 2001, Kärkäs et al. 2005).

This paper presents preliminary results of the first author's MSc project. We report field observations and chemical data for a frozen lake that is located in the vicinity of nunatak Basen, north Vestfjella, near the coast of western Dronning Maud Land (Fig. 1). We also provide compositional data for water samples from streams and ponds on the nunatak. We use the data to constrain the contribution of surface melting to the water budget during the austral summer of 2003–2004. The field observations and sampling were performed during the FINNARP (Finnish Antarctic Research Program) 2003–2004 expedition.

Background and study area

Frozen lakes have been reported to form in the coastal regions of Dronning Maud Land. They are mainly located in the vicinity of nunataks but they are also formed in nunatak-free areas (Bintanja 1999, Liston et al. 1999a, 1999b). The lakes are a couple of kilometers long and 0.5–1 km wide and are typically located in places where the thickness of the ice sheet exceeds 500 m (Liston et al. 1999b). In north Vestfjella, a 4 km long and 2 km wide blue ice field immediately southwest of nunatak Basen constitutes a profound local glacial feature. During the austral summer, a subsurface liquid layer develops in the blue ice field, which is hereafter designated as (frozen) Lake Basen. Water from the lake has been utilized by Finnish and Swedish Antarctic expeditions since 1988. The lake is bounded to the west by a snow ridge that extends from the nunatak some 2 km to southwest (Figs. 2 and 3). The eastern and southern limits of the blue ice area are less conspicuous due to gentle glacial topography and seasonal variations in the snow-cover. A 300 m long supraglacial till surfaces some 2 km from Basen and strikes southwest near the eastern margin of the lake (Figs. 2 and 3). The surface of the lake is only partially snow-free during the summer. Adjacent to the northern, western, and southwestern

Fig. 2. The areal extent of Lake Basen, north Vestfjella, western Dronning Maud Land. Also shown is the dominant wind direction (Photo: TERRA/ASTER © Canadian Space Agency, 1997).

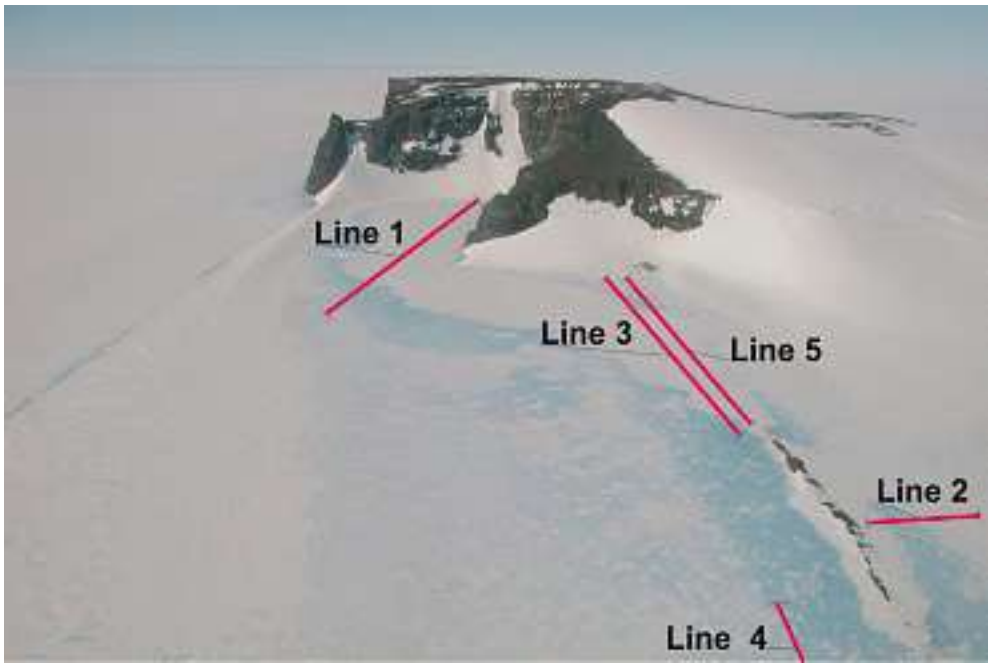


Fig. 3. Sampling transects in Lake Basen. View towards northeast. Transects 1 through 4 were sampled between 10.12.2003 and 3.1.2004, whereas transect 5 was sampled on 21.1.2004. Transect 4 extends southwest outside the field of the photo. The length of the supraglacial till (in the lower right corner) is 300 meters (Photo: E. Kärkäs, 2001)

bluffs of the nunatak, there is a moat-like narrow zone which develops into a narrow pool during the summer season (Fig. 4).

Objectives, materials and methods

Located adjacent to an unusually flat nunatak, Lake Basen was initially considered to be a promising area for addressing the influence of surface melting on the development of liquid layers in melt-induced blue ice areas. The main objectives of this study were to record 1) the surface morphology and 2) development of the liquid water layer in Lake Basen during the austral summer of 2003–2004, and 3) to establish the chemical composition and possible compositional variations of the water. Objectives 1 and 2 were approached by field observations, photography, and drilling during the period 15.11.2003–28.1.2004. For the chemical analyses, the first author collected water samples along four sampling transects (transect 1 on 10.12., transect 2 on 14.12., transect 3 on 19.12., and transect 4 on 3.1.) in different parts of the lake (Fig. 3). The sampling interval varied between 50 and 200 m. A new set of samples (transect 5) was collected along transect 3 on 21 January to monitor temporal variations. Additionally, water samples were collected from ponds and streams on the high plateau and on the slopes and cliffs of nunatak Basen in order to constrain the compositional characteristics of surface melts. Previous geological studies have reported salt deposits on bedrock and underneath rocks on the surface of the basal till on Basen (Lintinen & Nenonen 1997). Such precipitates were also sampled as reference material.

The water samples were collected according to the guidelines delineated for ITASE (International Trans-Antarctic Scientific Expedition) by Twickler & Whitlow (2005). The main cations (Na^+, K^+, Mg^{2+}, Ca^{2+}) and anions (Cl^-, NO_3^-, SO_4^{2-}) were analyzed using a Dionex ICS-1000 and a Dionex DX-120 ion chromatograph, respectively, at the Finnish Forest Research Institute, Rovaniemi. The analytical methods have been described by Virkkunen (2004). The mineral samples were identified using an X-ray diffractometer (Phillips X'pert, PW 1827/91) at the Geochemical Laboratory, University of Helsinki. The albedo measurements were performed with an EP 16 Pyrano-albedometer sensor, manufactured by Middleton, which covers the 300–3000 nm wavelengths.

Observations and results

At the beginning of the field season, from 15.–29.11., no liquid water was discovered in drill holes, but melting had commenced in the moat which was utilized by FINNARP as a source of household water (Petteri Järmälä and Petri Heinonen, personal communication, 2003). Albedo measurements yielded average values

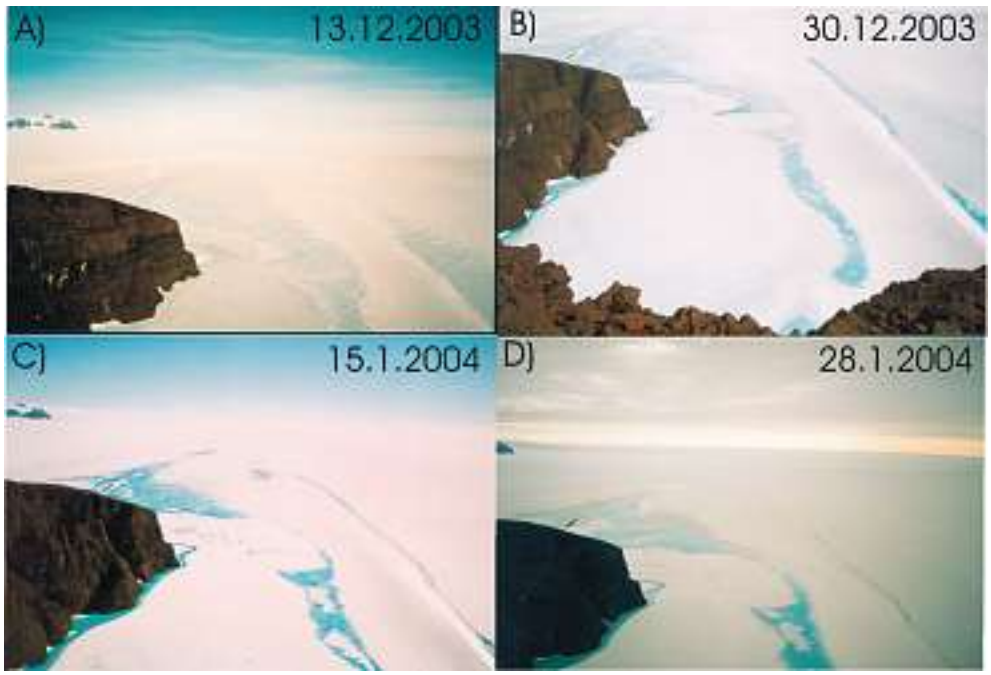


Fig. 4. Evolutionary stages of Lake Basen. A) In mid-December, percolating liquid water was observed in drill holes in the snow-free parts of the lake at depth of ca. 20–60 cm below the surface. B) At the end of December, a 15–40 cm thick liquid water layer was found from all over the lake. The ice cover was ca. 7–25 cm thick. Note the narrow mote-like feature at the base of the cliff. C) In mid-January, the lake had reached its maximum lateral extent. The streams from the nunatak were rapid. An open channel had formed in the central part of the lake. The flow of water towards southwest had formed a pond ca. 4 km away from the nunatak. The remaining ice cover in the snow-free parts was 5–7 cm and the liquid layer was ca. 60–120 cm thick. D) Freezing over of the channel had commenced. Streams were no longer rapid and the lake had frozen over. The liquid layer was still 60–120 cm thick. Sediments had been deposited from the current near the eastern margin of the lake (near the left edge of the image) (Photos: A. Lehtinen, 2003/2004)

of 0.44 for the blue ice area (31.12.2003–2.1.2004) and 0.76 for the adjacent snow-covered area (15.–18.1.2004). The latter value is slightly lower than the average value previously obtained by Kärkäs et al. (2002) for north Vestfjella and Heimefrontfjella (Fig. 1).

On December 1st, liquid water was observed in drill holes in the snow-free parts of the lake, for example, in the middle part of transect 2 (Fig. 3). The drill hole was filled with water which percolated through cracks and air bubbles in the ice. Water was not found in snow-covered parts of the lake.

In mid-December (13.12.2003), small streams ran from the nunatak to the lake and emptied into the moat. Percolating liquid water was found in several drill holes in snow-free locations around the lake (Fig. 4a).

A week later (19.12.2003), a distinctive 15–40 cm thick liquid layer was discovered in snow-free areas below a 7–25 cm thick ice surface. Melting of the surface ice was observed along the edges of columnar ice crystals.

By the end of December (30.12.2003), the areal extent and the thickness of the liquid layer had increased notably (Fig. 4b). The size and lateral geometry the liquid layer is reflected in Figure 4 by the deep blue tint of the ice surface (cf. Fig. 2). At this point, a liquid layer was also discovered in snow-covered parts and in the southwestern reaches of the lake.

By mid-January (15.12.2004), the lake attained its largest areal extent. In places there was no surface ice, and elsewhere in the snow-free areas the ice thickness was less than 5–7 cm. In snow-covered parts the thickness of the surface ice was 10–20 cm. The thickness of the liquid layer ranged from 60 to 130 cm. A current was observed in the open channel which ran southwest and generated a supraglacial pond some 4 km away from Basen. The melt water streams on the nunatak were rapid. A wide outlet, separated from the lake by the snow ridge, was developed and connected the western moat with the lake near the supraglacial till (Fig. 4c). The snow-covered parts and the shores of the lake became slush and sludge.

Five days later (20.1.2004), the streams had weakened, some of them dried up, and the open water in the central part of the lake froze over (Fig. 4d). In the marginal parts of the lake, dark-coloured sediments were deposited in subsurface channels. The southwesterly flow continued and parts of the lake ran dry and left behind large cavities beneath an undulating ice surface. By 28 January the streams had frozen and

Table 1. Chemical data for lake and stream water samples from Basen, north Vestfjella, western Dronning Maud Land.

Specimen number (type)	Distance to Basen (m)	Cl-	NO3-	SO42-	Na+	K+	Mg2+	Ca2+
BAV 3 (Line 1)	~1000	8.6	bdl.	27.9	8.7	0.1	4.4	185.8
BAV 37 (Line 3)	600	62.3	20.7	52.0	45.7	17.2	15.4	130.0
BAV 38 (Line 3)	500	49.1	bdl.	64.1	35.9	10.7	4.0	75.0
BAV 39 (Line 3)	100	124.6	bdl.	99.7	93.6	10.0	39.3	270.1
BAV 51 (Line 5)	600	445.3	11.3	316.3	268.4	11.4	79.3	517.8
BAV 52 (Line 5)	500	339.8	17.1	321.0	217.3	7.4	70.8	518.7
BAV 55 (Line 5)	100	323.2	bdl.	262.4	214.2	11.4	135.6	948.3
BAV 35 (Stream)		14649.6	889.4	16309.7	9549.1	123.6	5329.4	34253.6

Ion concentrations are given in parts per billion (ppb). Line 1 was collected on December 10, Line 3 on December 19, Line 5 on January 21, and the stream water specimen on December 17. Concentrations of NO3- in four specimens are below the detection limit (bdl.).

the channels and the pond froze over. The surface ice in the central part of the lake was measured to be ca. 15 cm thick, while the liquid layer under the ice surface was still 60–120 cm.

X-ray diffractometer analyses showed the precipitate samples to be $CaCO_3$ (aragonite and calcite) and Na_2SO_4 (thenardite). Chemical compositions of representative water samples are listed in Table 1. The prevalent ions in the lake samples are Ca^{2+} (<950 ppb), Cl^- (<450 ppb), and SO_4^{2-} (< 320 ppb). Sample BAV 3 from transect 1 was collected ca. 1 km away from Basen on 10 December and is considered to represent incipient subsurface melting. The overall ion concentrations in this sample are markedly low

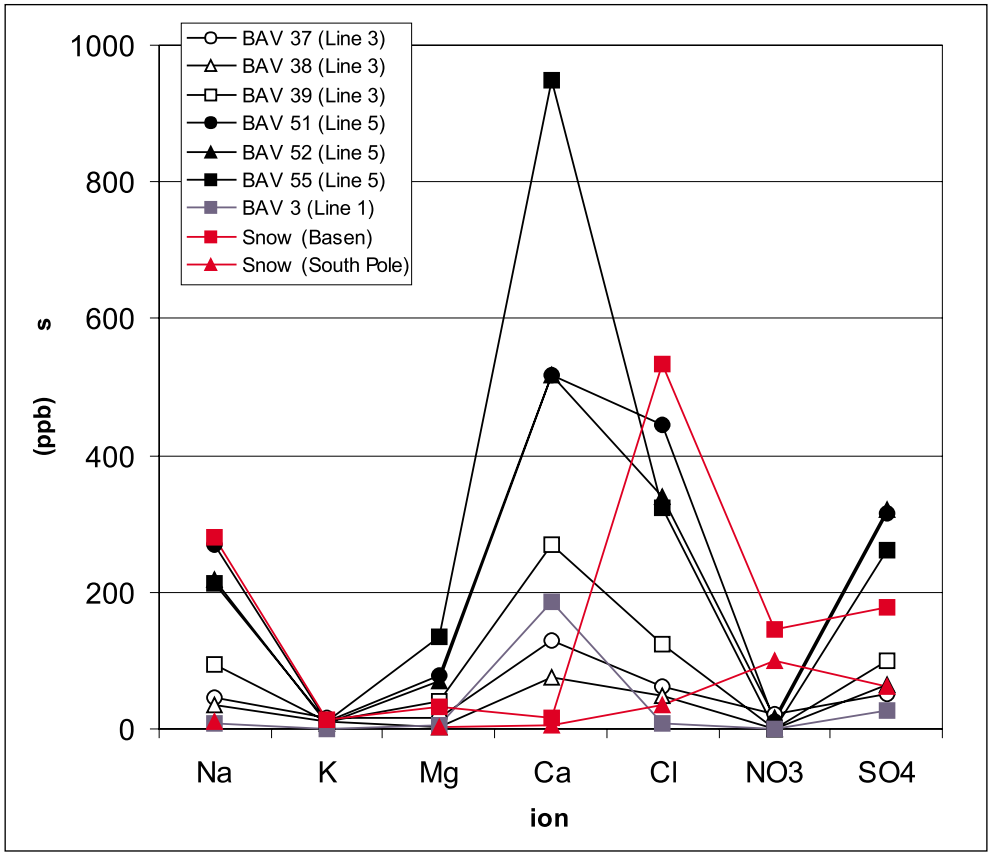


Fig. 5. Chemical comparison of water and snow. Sample BAV 3 from transect 1 represents lake water on 10.12.2003; samples BAV 37, BAV 38 and BAV 39 from transect 3 represent lake water on 19.12.2003; and samples BAV 51, BAV 52 and BAV 55 from transect 5 represent lake water on 21.1.2004. The average composition of snow from north Vestfjella (Basen) and from the East Antarctic plateau (South Pole) are from Kärkäs et al. (2005) and Whitlow et al. (1992), respectively.

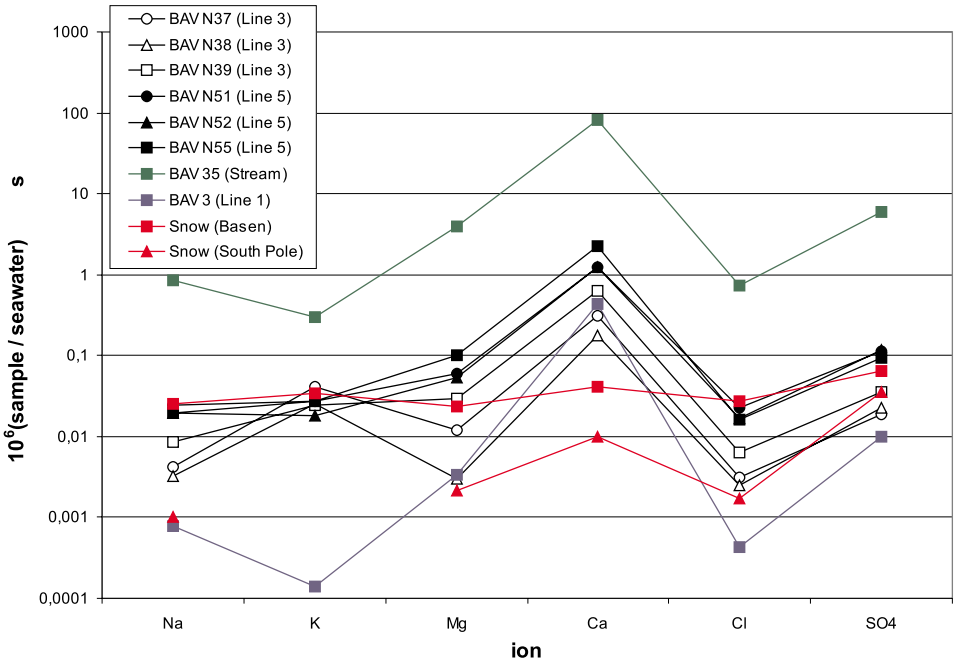


Fig. 6. Seawater-normalized chemical fingerprints for lake water, stream water and snow. Sample descriptions are presented in Fig. 5. Sample BAV 35 represents stream water on 17.12.2003. Normalization values are from Riley & Chester (1971). Data sources as in Figure 5.

(typically <30 ppb), and, apart from Ca^{2+} and NO_3^-, the composition resembles that of snow from the East Antarctic plateau (Fig. 5). Similarly, samples BAV 37, BAV 38, and BAV 39, which were collected along transect 3 on 19.12., have low ion contents and their overall composition is clearly different from that of snow samples from the study area (Kärkäs et al. 2005). Specifically, the concentrations of Na^+, Cl^-, NO_3^-, and SO_4^{2-} are clearly lower, whereas the concentrations of Ca^{2+} are markedly higher in the lake samples (Fig. 5). Samples BAV 51, BAV 52 and BAV 55 were collected along transect 5 on January 21. The highest Ca^{2+} and Mg^{2+} concentrations in transects 3 and 5 were measured in samples that were collected closest to the nunatak (BAV 39 and BAV 55) and the compositional ranges for both transects are notably reduced when these samples are excluded (Table 1). Comparison of samples from transects 3 and 5 indicates a marked 4-fold increase in the averaged ion concentrations during the sampling interval. The concentrations of K^+ and NO_3^- do not increase. Instead, they show low, but variable, concentrations at all times. Ion concentrations in the ponds and streams are markedly variable (e.g. Ca^{2+} 1220–34250 ppb) but consistently high and frequently over 5000 ppb (Table 1). The sample in Table 1 was collected on 17 December and records the highest ion concentrations in our stream water samples.

Figure 6 shows a comparison between lake, stream and snow samples, and average seawater. Snow from the coastal region near Basen has a flat seawater-normalized fingerprint, and, accordingly, near-seawater ion ratios. By comparison, snow from the East Antarctic plateau (South Pole) has clearly lower ion concentrations, notably higher SO_4^{2-}/Cl^- and somewhat higher Ca^{2+}/Mg^{2+}. The Figure indicates a similarity between lake water, stream water, snow, and seawater in terms of Cl^-/Na^+ which is always within the range 1.4–1.9, i.e. similar to the corresponding ratio for seawater (1.8; Riley & Chester, 1971). Otherwise, all water samples can be distinguished based on their spiked chemical fingerprints. The lake and stream samples are most clearly differentiated from snow based on their high Ca^{2+}/Mg^{2+}, which is approximately 20 times higher than in the seawater of Riley & Chester (1971). Generalizing, the ion ratios in the lake water remain rather constant during the season. For example, Ca^{2+}/Mg^{2+} is 7–8 on 19.12. and ca. 7 on 21.1.; one sample of the 19.12. dataset is anomalous with Ca^{2+}/Mg^{2+} of ca. 19.

Discussion and conclusions

Chemical analyses of streams and ponds from nunatak Basen showed them to be highly saline compared with snow from the same region. In addition to high overall ion concentrations, the chemical fingerprint of the stream water differs from that of snow by high Na^+/K^+, Mg^{2+}/K^+, SO_4^{2-}/Cl^- and, specifically, Ca^{2+}/Mg^{2+}

(Fig. 6). These characteristics facilitate the detection of even a small quantity of stream water in Lake Basen. These key differences between the stream water and snow most probably reflect the involvement of precipitates of calcium and magnesium carbonate and thenardite from the nunatak. The principal source of most ions in Antarctic snow and ice is the ocean (Gjessing 1984, Mulvaney & Wolff 1994, Legrand & Mayewski 1997). In Lake Basen, however, the principal source of Ca and Mg is probably amygdule-hosted carbonate in the Jurassic basalts which make up the bedrock (Kallio 2003). The decoupling of K^+ and NO_3^- from other ions, i.e. their consistently low concentrations, is readily associated with algae, lichen, and moss growths that have been observed in the ponds and streams on the nunatak. Magnesium is also used up by the photosynthetic organisms, but the effect is overprinted by the release of Mg from the mineral precipitates.

The pronounced increase in ion concentrations during the summer and the compositional similarity between the stream water and lake water demonstrates that surface melting has a strong influence on the liquid layer in Lake Basen. The low albedo of the blue ice area (0.44) compared with adjacent snow-covered areas (0.76) and the notably low concentrations of ions in the first water samples indicate that the initial melt is produced by subsurface melting, i.e. the solid state greenhouse effect (Brandt & Warren 1993). We conclude that during the austral summer of 2003 – 2004, Lake Basen evolved as an open system and the liquid layer resulted from combined subsurface and surface melting.

One of the pertinent questions with regard to the evolutionary cycle of Lake Basen arises from the low ion concentrations in the first lake water samples. Given that the last water samples of our study represent a thick liquid layer of up to 120 cm, the late-stage evolution of the lake remains poorly constrained. However, we consider that the observed increase in ion concentrations is probably an annual phenomenon in this lake. Hence, a dramatic compositional change during the fall is required to explain the low ion concentrations in the incipient melt. Our field observations indicate that surface melting already diminished in mid-January. Subsurface melting probably continues for some time after the influx of the surface melting component has ceased to be significant and dilutes the ion concentrations in the liquid layer as its volume decreases due to freezing and outflow. However, the compositional differences between the first melt and Antarctic snow from the coastal area and from the East Antarctic Plateau (Fig. 6) imply that a diluted surface water component remains in the ice.

The high ion concentrations and, specifically, the distinctive chemical fingerprint of the stream water render it an easily detectable component. Based on ion concentrations in the initial melt and the stream water (e.g. Mg), we estimate that the maximum proportion of stream water in the liquid layer may be ≤2 wt.%, although the highly variable compositions undermine mass balance calculations. Our study casts new light on the evolution of frozen lakes in Antarctica. Our results may be applicable to other near-nunatak frozen lakes, although the example from Basen may represent an epitome of surface melt-influenced systems due to the unusually flat topography of the nunatak.

Acknowledgements

The authors thank Kristiina Virkkula for chemical analyses, Olli-Pekka Mattila for cooperation with albedo measurements, and Martti Lehtinen for X-ray diffractometer analyses of mineral precipitates. Discussions with Eija Kanto (née Kärkäs) on snow chemistry were elucidative. Mikko Haaramo helped us to recognize the influence of photosynthetic organisms on ion species. Kimmo Kanto kindly gave us the permission to use Terra aster images. FINNARP logistics personnel gave extensive help during the fieldwork. The reviews of Matti Eronen and Matti Saarnisto helped to improve the manuscript. Financial support was provided by the Academy of Finland (grant # 54087).

REFERENCES

Bintanja, R. 1999. On the glaciological, meteorological and climatological significance of Antarctic blue ice areas. Reviews of Geophysics 37, 337–359.

Brandt, R. & Warren, S. 1993. Solar-heating rates and temperature profiles in Antarctic snow and ice. Journal of Glaciology 39, 99–110.

Bøggild, C.E., Winther, J.-G., Sand, K. & Elvehøy, H. 1995. Sub-surface melting in blue-ice fields in Dronning Maud Land, Antarctica: observations and modelling. Annals of Glaciology 21, 162–168.

Echelmeyer, K., Clarke, T.S. & Harrison, W.D. 1991. Surficial glaciology of Jakobshavns Isbrae, West Greenland: Part I. Surface morphology. Journal of Glaciology 37, 368–382.

Gjessing, Y. 1984. Marine and non-marine contribution to the chemical composition of snow at the Riiser-Larsenisen ice shelf in Antarctica. Atmospheric Environment 18, 825–830.

Isaksson, E., Karlén, W., Mayewski, P., Twickler, M. & Withlow, S. 2001. A high-altitude snow chemistry record from Amundsenisen, Dronning Maud Land, Antarctica. Journal of Glaciology 47, 489–496.

Kallio, A. 2003. Low-grade metamorphism of a Jurassic continental flood basalt flow in Vestfjella, Dronning Maud Land, Antarctica. University of Helsinki, Department of Geology, Unpublished M.Sc. Thesis, 96 p.

Kärkäs, E., Granberg, H.B., Kanto, K., Rasmus, K., Lavoie, C. & Leppäranta, M. 2002. Physical properties of the seasonal snow cover in Dronning Maud Land, East Antarctica. Annals of Glaciology 34, 89–94.

Kärkäs, E., Teinilä, K., Virkkula, A. & Aurela, M. 2005. Spatial Variations of surface snow chemistry during two austral summers in western Dronning Maud Land, Antarctica. Atmospheric Environment 39, 1405–1416.

Legrand, M. & Mayewski, P. 1997. Glaciochemistry of Polar Ice Cores: A Review. Reviews of Geophysics 35, 219–243.

Lintinen, P. & Nenonen, J. 1997. Water-soluble chemistry and weathering characteristics of some tills in Western Dronning Maud Land, Antarctica. Bulletin of the Geological Society of Finland 69, 57–71.

Liston, G.E., Bruland, O., Winther, J.-G., Elvehøy, H. & Sand, K. 1999a. Meltwater production in Antarctic blue-ice areas: sensitivity to changes in atmospheric forcing. Polar Research 18, 283–290.

Liston, G.E., Winther, J.-G., Bruland, O., Hallgeir, E. & Sand, K. 1999b. Below-surface ice melt on the coastal Antarctic ice sheet. Journal of Glaciology 45, 273–285.

Mulvaney, R. & Wolff, E.W. 1994. Spatial variability of the major chemistry of the Antarctic ice sheet. Annals of Glaciology 20, 440–447.

Riley, J.P. & Chester, R. 1971. Introduction to Marine Chemistry. London: Academic Press, 465 p.

Stenberg, M., Isaksson, E., Hansson, M., Karlén, W., Mayewski, P., Twickler, M., Withlow, S. & Gunderstrup, N. 1998. Spatial variability of snow chemistry in western Dronning Maud Land, Antarctica. Annals of Glaciology 27, 378–384.

Twickler, M. & Whitlow, S. 2005. Appendix B, Guide for the Collection And Analysis Of ITASE Snow And Firn Samples. /ITASE (International Trans-Antarctic Scientific Expedition), 200 Years of Past Antarctic Climate and Environmental Change, Science and Implementation Plan. Available from http://www.ume.maine.edu/itase/sci_plan/appb.html.

Virkkunen, K. 2004. Snowpit studies in 2001–2002 in Lomonosovfonna, Svalbard., University of Oulu, Department of Chemistry, Unpublished M.Sc. Thesis, 73 p.

Winther, J.-G., Elvehøy, H., Bøggild, C.E., Sand, K. & Gliston, G. 1996. Melting, runoff and the formation of frozen lakes in a mixed snow and blue-ice field in Dronning Maud Land, Antarctica. Journal of Glaciology 42, 271–278.

Tätä julkaisua myy

GEOLOGIAN TUTKIMUSKESKUS (GTK)
Julkaisumyynti
PL 96
02151 Espoo
☎ 020 550 11
Telekopio: 020 550 12

GTK, Kuopion yksikkö
Kirjasto
PL 1237
70211 Kuopio
☎ 020 550 11
Telekopio: 020 550 13

GTK, Rovaniemen yksikkö
Kirjasto
PL 77
96101 Rovaniemi
☎ 020 550 11
Telekopio: 020 550 14

E-mail: julkaisumyynti@gtk.fi
WWW-address: www.gsf.fi

Denna publikation säljes av

GEOLOGISKA FORSKNINGSCENTRALEN (GTK)
Publikationsförsäljning
PB 96
02151 Esbo
☎ 020 550 11
Telefax: 020 550 12

GTK, Kuopio enhet
Biblioteket
PB 1237
70211 Kuopio
☎ 020 550 11
Telefax: 020 550 13

GTK, Rovaniemi enhet
Biblioteket
PB 77
96101 Rovaniemi
☎ 020 550 11
Telefax: 020 550 14

ISBN 951-690-930-2
ISSN 0782-8535

This publication can be obtained from

GEOLOGICAL SURVEY OF FINLAND (GTK)
Publication sales
P.O. Box 96
FIN-02151 Espoo, Finland
☎ +358 20 550 11
Telefax: +358 20 550 12

GTK, Kuopio Unit
Library
P.O. Box 1237
FIN-70211 Kuopio, Finland
☎ +358 20 550 11
Telefax: +358 20 550 13

GTK, Rovaniemi Unit
Library
P.O. Box 77
FIN-96101 Rovaniemi, Finland
☎ +358 20 550 11
Telefax: +358 20 550 14

9 789516 909304