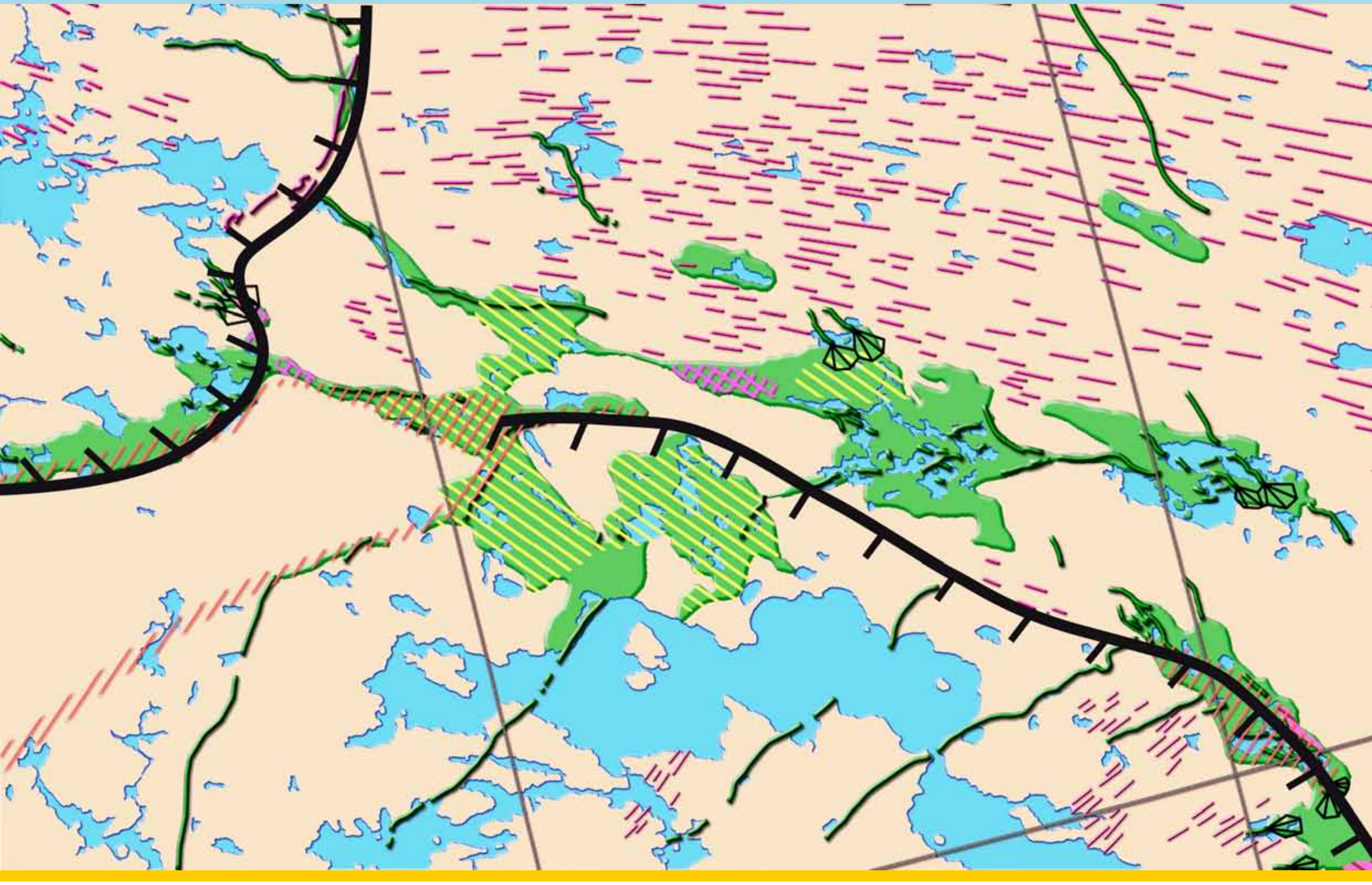
Late Weichselian deglaciation chronology and palaeoenvironments in northern Karelia, NW Russia

Niko Putkinen



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LATE WEICHSELIAN DEGLACIATION CHRONOLOGY AND PALAEOENVIRONMENTS IN NORTHERN KARELIA, NW RUSSIA

by

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ACADEMIC DISSERTATION
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Cover: Glacial geomorphology and the end moraine zones of the Kalevala area.
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ABSTRACT

The Scandinavian Ice Sheet (SIS) was one of the largest ice sheets in Eurasia during the Weichselian glaciation. It attained its maximum extent in the east during the Late Weichselian between 18 000–15 000 years ago, covering the whole of Fennoscandia, northwestern Russia and northern Continental Europe and coalescing with the Barents Ice Sheet and the British Ice Sheet. The Weichselian deglaciation was interrupted by a sudden climate cooling episode known as the Younger Dryas Stadial (ca. 12 800–11 500 years ago). This thesis examines the palaeoenvironments in front of the SIS during the Late Weichselian and Early Holocene (ca. 12 500–11 000 years ago). The eastern and western Rukajärvi, Kalevala and Pääjärvi end moraines were dated in order to reconstruct the position of the ice margin in Russian Karelia during the last deglaciation.

A number of palaeoenvironmental research techniques were used. Morphologically different land systems in the field were examined by glacial geomorphological studies and in aerial photos and satellite images. Internal structures of glaciofluvial formations, especially delta plains and beach deposits, were studied using ground penetrating radar (GPR). These flat plains were used as elevation data to produce distance diagrams for GIS-based reconstruction of the White Sea Basin water levels in the Younger Dryas Stadial. Early development of the White Sea was also studied, using conventional shoreline displacement methods. Finally, deglaciation chronology of the study area was determined using a combination of glacial varve chronology and palaeomagnetic dating of three lacustrine sediment sequences from basins situated in the end moraine zones. Geomorphology, ^{14}C AMS dating and previous knowledge of ice lobe retreat rates were also used for chronological analysis.

The results indicate that the Kuusamo-White Sea and Northern Karelian ice lobes of the SIS operated in the area and deposited several fan-shaped geomorphological landform patterns and end moraine zones. Indications were found that four ice advances occurred in the area during the Late Weichselian and Early Holocene deglaciation. The eastern arc of the Rukajärvi end moraine, in the Rukajärvi-Belomorsk area, marks the southernmost limit of the Kuusamo-White Sea ice lobe advance, the oldest in the area. The Onega sub-ice lobe (part of the Kuusamo-White Sea ice lobe) apparently advanced from the north-northwest to the eastern Rukajärvi end moraine zone, which was deposited ca. 12 500–12 300 years ago. As deglaciation continued, the northern Karelian ice lobe was formed and advanced from west to east, depositing the western arc of the Rukajärvi end moraine 11 700–11 600 years ago.

The North Karelian ice lobe retreated westward but advanced again and the Kalevala end moraine was deposited in front of the ice margin 11 400–11 300 years ago. Further north, the Pääjärvi end moraine was accumulated at the margin of the Kuusamo-White Sea ice lobe 11 000 years ago. There were indications that the western Rukajärvi end moraine was correlated in time with the Salpausselkä II in southeastern Finland, while the Kalevala end moraine was possibly deposited as the same time as Salpausselkä III in southwestern Finland.

The results also suggest that the Kuusamo-White Sea ice lobe terminated in a glacial lake that occupied the White Sea Basin and adjacent land areas. Around 12 050 years ago the waters of this ice lake burst into the Barents Sea via the Gorlo Strait and the lake water level fell by approximately 50–60 metres within 200 years. After this event the White Sea Basin was connected to the

Barents Sea. The northern parts of the Kuittijärvi and Tuoppajärvi sub-ice lobes terminated in shallow water.

Keywords (GeoRef Thesaurus, AGI): glacial geology, Scandinavian ice sheet, deglaciation, ice-marginal features, end moraines, chronology, paleoenvironment, glacial lakes, Weichselian, Younger Dryas, Russian Federation, Republic of Karelia, Kalevala, Belomorsk, White Sea

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LIST OF ORIGINAL PAPERS

This thesis is based on the following articles, which will be referred to in the text by their Roman numerals.

I Putkinen, N. & Lunkka, J. P. 2008. Ice stream behaviour and deglaciation of the Scandivian Ice Sheet in the Kuittijärvi area, Russian Karelia. Bulletin of the Geological Society of Finland 80 (1), 19–37.

II Lunkka, J. P., Putkinen, N. & Miettinen. A. Shoreline displacement in the Belomorsk area, NW Russia during the Younger Dryas Stadial. Manuscript

III Pasanen, A., Lunkka, J. P. & Putkinen, N. 2010. Reconstruction of the White Sea Basin during the late Younger Dryas. Boreas 39, 273–285.

IV Putkinen, N., Lunkka, J. P., Ojala, A. E. K. & Kosonen, E. Deglaciation history and age estimate of the Younger Dryas end moraines in the Kalevala region, NW Russia. Manuscript

1 INTRODUCTION

The Scandinavian Ice Sheet (SIS) was one of the largest ice sheets in Eurasia during the Weichselian glaciation (Svendsen et al. 2004). It attained its maximum extent during the Late Weichselian period, covering the whole of Fennoscandia, northwestern Russia and northern Continental Europe, and coalescing with the Barents Ice Sheet and the British Ice Sheet. The southwestern part of the ice sheet was at its maximum 22 000–20 000 years ago (all ages used in this synopsis, except those are presented in preview of the original papers, are calendar years before present), while in the east it reached its maximum extent 18 000–15 000 years ago (Fig. 1) (cf. Ehlers et al. 2004, Larsen et al. 1999, Boulton et al. 2001, Lunkka et al. 2001, Johansson et al. 2011, Svendsen et al. 2004). On the eastern and southeastern flank of the SIS, the ice retreat from its maximum position is thought to have been a more or less continuous process. However, prominent end moraine complexes such as the Pommerian (ca. 14 600 years ago) and Gardno (ca. 15 400 years ago) in the southeast, and the Vepsian-Krestets (ca. 15 000 years ago), Luga (ca. 14 000 years ago) and Neva (ca. 13 000 years ago) in the east-southeast were deposited during minor standstills and oscillations of the ice margin during the course of deglaciation (cf. Kozarski 1986, Ekman & Iljin 1991, Boulton et al. 2001, Svendsen et al. 2004, Rinterknecht et al. 2006).

During the Younger Dryas Stadial (ca. 12 800–11 500 years ago; cf. Muscheler et al. 2008) glaciers advanced worldwide, not only in Fennoscandia but also in Scotland, the Americas and the high mountain areas of the Alps, Asia and New Zealand as a result of climate cooling. The Salpausselkäs and their relatives that run around the whole of Fennoscandia and northwestern Russia were formed at the ice margin during that time. These moraines represent the most prominent glacial geomorphological feature in Fennoscandia and NW Russia, marking the Younger Dryas Stadial ice sheet margin (cf. Donner 1995, Rainio et al. 1995). The asynchronous SIS marginal formations that formed in Fennoscandia during the Younger Dryas Stadial can be traced over a distance of more than 2500 kilometres.

The Younger Dryas Stadial ice margin, which is marked by the Salpausselkäs, Tuupovaara, Koitere and Pielisjärvi end moraines in Finland and the Rukajärvi, Kalevala and Pääjärvi and Keiva end moraines in northwestern Russia (Fig. 1), can be traced from Finland to Russian Karelia and on to the Kola Peninsula (Ekman & Iljin 1991, Rainio et al. 1995, Lunkka et al. 2004, Johansson et al. 2011). In the northern, western and southern parts of Fennoscandia, the Younger Dryas Stadial ice margin followed the Lyngen, Tromsø, Tautra, Herdla and Ra moraines in Norway and the Central Sweden end moraines (Skövde and Billingen moraines) in Sweden (Fig. 1) (Lundqvist & Wohlfarth 2001, Mangerud 2004). In the southern and western parts of the SIS, the chronology of the Younger Dryas Stadial end moraines is largely based on ^{14}C dating, biostratigraphy and varve chronologies (cf. Sauramo 1923, Lundqvist & Wohlfarth 2001, Mangerud 2004). Lundqvist & Wohlfarth (2001) were of the opinion that the Swedish Skövde and Billingen moraines were formed during the Younger Dryas Stadial. Both of these end moraines were deposited into the Baltic Ice Lake.

In Norway, the Central Sweden Skövde-Billingen end moraines continued as the Ra end moraine in the Oslofjorden area towards the southwest (Sørensen 1979, Mangerud 2004). In the Oslofjorden area the Ås and Ski end moraines were deposited at the end of the Younger Dryas Stadial (Mangerud 2004). Further north, the Halsnøy end moraine at the mouth of the Hardagerfjord and also the Herdla end moraine in the Bergen area are correlated to the Ski moraine in Oslofjorden. There are no Younger Dryas end moraines between Ålfoten and Norrland, but in

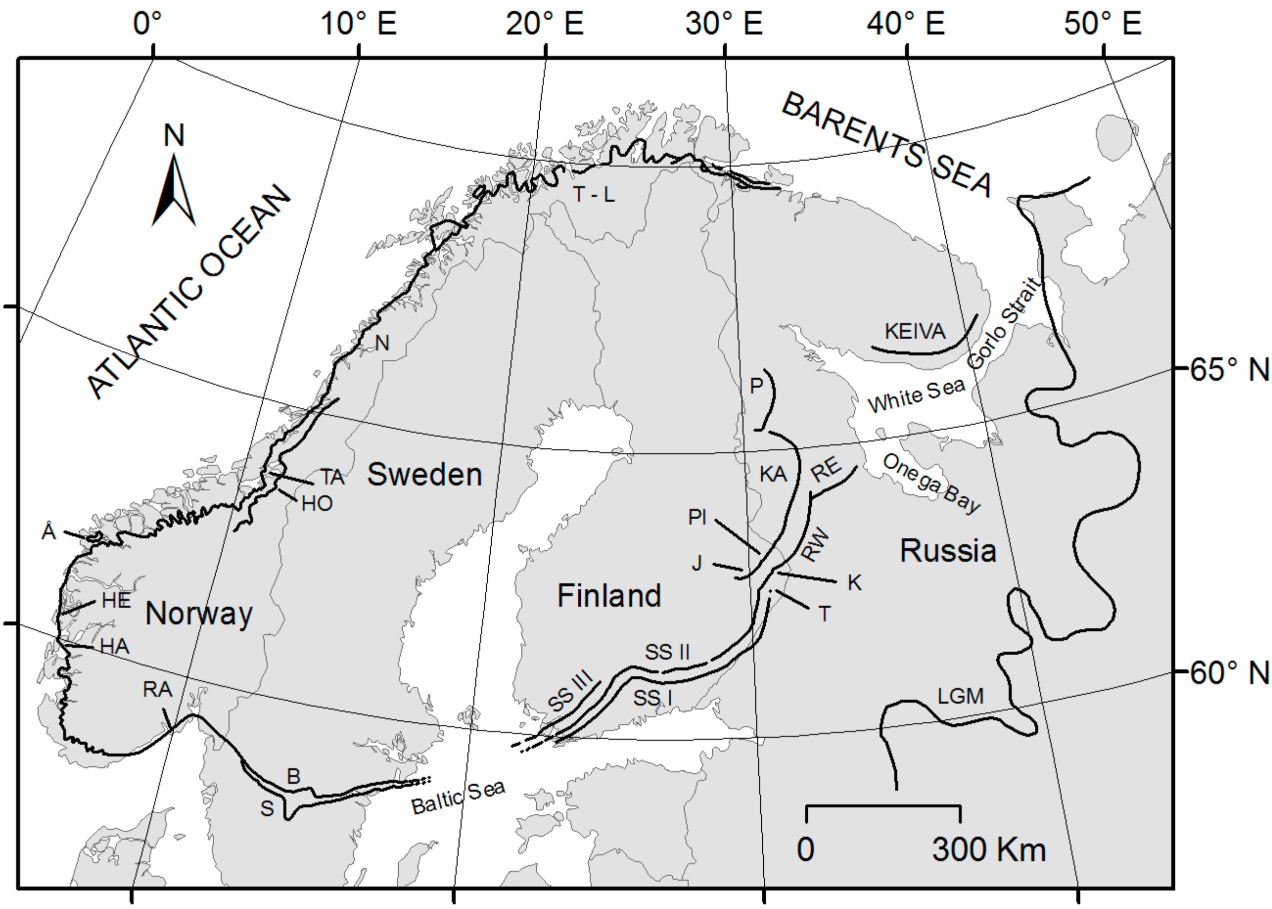

Fig 1. Younger Dryas Stadial end moraines and their relatives in Fennoscandia, modified after Mangerud (2004). The main end moraines in the NW Russia are: KA = Kalevala, P = Pääjärvi, RE = Eastern Rukajärvi, RW = Western Rukajärvi, K = Keiva. The main end moraines in southern Finland are: SS I-III = First, second and third Salpausselkäs, PL = Pielisjärvi, K = Koitere, T = Tuupovaara, J = Jaamankangas ice marginal deposit. The main end moraines in Scandinavia are: S = Skövde, B = Billingen, RA = Ra, HA = Halsnøy, HE = Herdla, Å = Ålfoten, HO = Hoklingen, TA = Tautra, N = Norrland, T-L = Tromsø-Lyngen. LGM indicates the maximum extent of the Scandinavian Ice Sheet (SIS) on its eastern flank, modified after Demidov et al. (2004).

the Trondheim area, the Tautra and Hoklingen end moraines were formed ca. 12 500 and 12 000 years ago, respectively (Mangerud 2004). The Tautra end moraine continues towards the north to the Tromsø-Lyngen area, while the Hoklingen end moraine can be traced as far as southern Norrland (Sveian & Solli 1997). Norwegian end moraines were typically deposited onto the continental shelf.

The Salpausselkä I and II end moraines run across southern Finland (Fig. 1). They are continuous ridges, ca. 25 kilometres apart, that can be traced from the southern coast of Finland to the Finnish Karelia in southeastern Finland (Fig. 1) (Niemelä et al. 1993, Rainio 1996). In Finnish North Karelia, the Tuupovaara, Koitere and Pielisjärvi end moraines are discontinuous moraine zones that continue to Russian Karelia (Fig. 1). The existing chronology of the Salpausselkä and Pielisjärvi end moraines is based on varve chronology and biostratigraphical evidence, supported by radiocarbon ages constructed from the Salpausselkä area. The Finnish varve chronology (Sauramo 1918, 1923, 1926) originally covered 2210 varve years and was later corrected to include 2900 years (cf. Taipale & Saarnisto 1991). The chronology is based on the correlation of marker horizons formed as a result of water level changes in the Baltic Basin. The zero year, a thick varve horizon in the varved clay sequence, is generally thought to have been formed when the Baltic ice lake level (BIII) dropped 26–28 metres to the Yoldia Sea level (YI). At the same time or immediately afterwards, the Salpausselkä II end moraine was formed (Sauramo 1923, 1929, Donner 1995, Lunkka et al. 2004, Johansson et al. 2011). Using Finnish varve chronology, fixed to the zero year, together with ^{14}C AMS and cosmogenic exposure dating results (Saarnisto & Saarinen 2001, Tschudi et al. 2001, Rinterknecht et al. 2006), it has been concluded that Salpausselkä I was formed 12 300–12 100 calendar years ago and Salpausselkä II 11 800–11 600 years ago (cf. Lunkka et al. 2004, Donner 2011, Johansson et al. 2011).

The Rukajärvi end moraine in Russian Karelia (Fig. 1) runs 30 kilometres east of the southern stretch of the Kalevala end moraine from the Finnish Karelian Koitere end moraine to the Tiksha area (Fig. 3). The Rukajärvi end moraine was deposited in front of the North Karelian ice lobe and the Lake Onega sub-ice lobe (Figs. 1 and 2). According to Saarnisto & Saarinen (2001), the Rukajärvi end moraine was deposited 12 300–12 100 years ago.

Two other end moraines in Russian Karelia, the Kalevala and Pääjärvi end moraines, are relatively well mapped but they have not been directly dated. The Kalevala end moraine extends for 400 km from Jaamankangas in Finland to Kalevala in Russian Karelia, while the Pääjärvi end moraine runs from Kalevala to the northern part of Lake Pääjärvi (Figs. 1 and 2). The age estimates of these end moraines are based on radiocarbon dating of lake sediments, consideration of ice retreat rates and the correlation of the Russian Karelian end moraines to the Salpausselkäs in Finland (Ekman & Iljin 1991, Rainio et al. 1995, Boulton et al. 2001, Svendsen et al. 2004). It is generally accepted that the Kalevala end moraine was formed at the margin of the North Karelian lobe 11 800–11 600 years ago (Fig. 2) (Boulton et al. 2001, Svendsen et al. 2004) and that the Pääjärvi end moraine was formed in front of the Kuusamo-White Sea ice lobe 11 800–11 600 years ago (Boulton et al. 2001, Svendsen et al. 2004). However, Ekman & Iljin (1991) suggested that these end moraines were deposited at the same time as Salpausselkä I.

As outlined above, the absolute ages of the Younger Dryas Stadial moraines in Finland, Sweden and Norway are relatively well constrained, as are the palaeoenvironments that existed in front of the ice margin (Lundqvist 2004, Mangerud 2004, Lunkka et al. 2004, Rinterknecht et al. 2006). However, the absolute chronology of the last deglaciation and the palaeoenvironmental setting during the course of the ice retreat in Russian Karelia is not well known. The purposes of this thesis were therefore to reconstruct the palaeoenvironments in front of the SIS in Russian Karelia during the Late Weichselian and Early Holocene periods and to date the Rukajärvi, Kalevala and Pääjärvi end moraines. The overall aim was to reconstruct different positions of the ice margin during the Weichselian deglaciation.

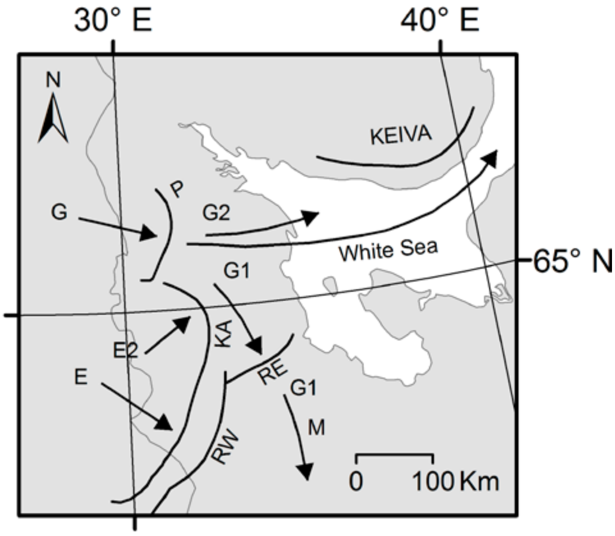

Fig 2. Ice streams in northern Karelia modified after Punkari (1985) (E = North Karelian ice lobe, G = Kuusamo-White Sea ice lobe, G1 = Onega sub-ice lobe, G2 = Tuoppajärvi sub-ice lobe and E2 = Kuittijärvi sub-ice lobe). End moraines and Maaselkä drainage divide (M) are also indicated, for abbreviations see Fig. 1.

2 STUDY AREA

The study area is located in Northern Karelia, on the western side of the White Sea in Russian Karelia. The area examined was 400 km long and 200 km wide, situated between the Maaselkä drainage divide and Lake Tuoppajärvi area (Figs. 2 and 3). Russian Karelia belongs to the eastern part of the Fennoscandian Shield (cf. Koistinen et al. 2001). The bedrock of the area consists of eroded Archaean basement metamorphic rocks and Paleoproterozoic volcanic and/or metasedimentary rocks with ultra-mafic intrusions (Eilu et al. 2008). Morphologically, this undulating area slopes towards the White Sea graben. The coastal area adjacent to the White Sea is flat, while the morphology of the inland area undulates to the amount of tens of metres depending on rock types. The altitude in the northwestern part of the study area ranges from 110 to 200 metres above the present day sea level (a.s.l.), while the altitude of the highest hills is 250–300 m a.s.l. In the western part the altitude ranges from 60 to 100 m a.s.l., the highest hills locally being 240–348 m a.s.l., while in the southwest it ranges between 120 and 145 m a.s.l., the highest point locally being 190 m a.s.l.

Three ice marginal zones and an interlobate complex consisting of glaciofluvial and/or till material and also streamlined moraines are the main glacial geomorphological landforms in the study area (cf. Ekman & Iljin 1991, Niemelä et al. 1993, Putkinen & Lunkka 2008, Elina et al. 2010). Otherwise, the Quaternary deposits that cover the undulating bedrock topography are relatively thin and consist of basal till. The lowland areas are covered by silt, clay and peat deposits (cf. Niemelä et al. 1993).

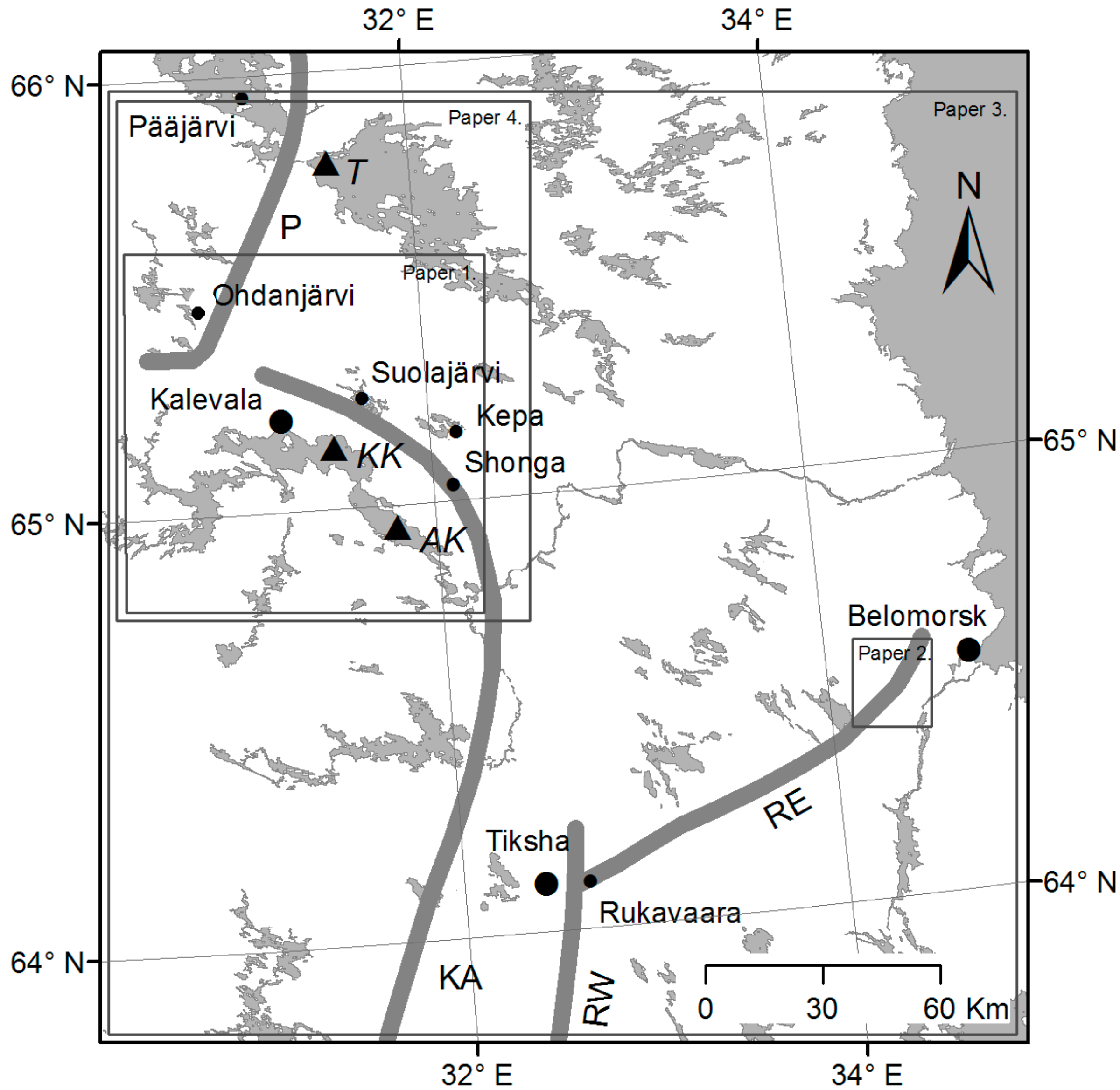

Fig 3. Main locations of the study area. Coring sites (triangles) are: AK = Ala-Kuittijärvi, KK = Keski-Kuittijärvi, T = Tuoppajärvi. End moraines (codes RE, RW, KA and P) and areas of interest in Papers I–IV are also indicated. For abbreviations see Fig. 1.

3 METHODS

The methods used in this work included a number of techniques applicable to palaeoenvironmental research in glacial environments. The work began with geomorphological mapping in the Karelian ice marginal zones. Aerial photography studies were conducted at the Institute of Geology, Karelian Research Centre, Petrozavodsk. Later, the glacial landscape of the study area was also mapped using satellite imagery according to the procedures outlined in Paper I. Aerial interpretations of the glacial landforms in the study area were checked during 10 field expeditions in the period 2002–2008. Particular attention was paid to glacial geomorphological formations deposited at the margin of the local ice lobes, ice flow directions and features indicating former water levels.

Glacial geomorphological studies concentrated on identifying different land systems that are morphologically recognisable in the field and in aerial photographs and satellite images (cf. Punkari 1985). These glacial landforms comprise characteristic landform associations that can be linked to glacier retreat history (cf. Stokes & Clark 2001). Therefore, mapping of landforms such as drumlins, drumlinoids and megaflutes, non-streamlined topography, eskers and ice marginal landforms was used to reconstruct palaeo-ice lobes (Paper I) (cf. Stokes & Clark 2003). Paper I utilised a modified land-system approach to reveal the ice lobe pattern in the study area.

Internal structures of glaciofluvial formations, especially flat plains, were studied using ground penetrating radar (GPR) (Paper III) and also from shallow cable pits and ditches and shallow test pits (Papers I and III). GPR is a powerful, non-destructive electromagnetic profiling tool for revealing sedimentary structures in shallow subsurface sediment sequences. The method is based on reflection profiling of pulsating electromagnetic energy two-way travel times between a surface and electrical boundaries, which typically exist at sediment interfaces with different dielectrical properties depending on water content and lithology (cf. Annan & Davis 1976, Daniels et al. 1988). In Papers II and IV, a study of soft sediment cores was carried out using conventional sedimentological logging and description methods as outlined in these papers (cf. Håkanson & Jansson 1983).

The elevation data collected from the delta plains and associated beach deposits were used to estimate former water levels and to reconstruct distance diagrams, which are commonly used for determination of ancient shoreline gradients. In Paper III the shoreline gradient in the east of the Kalevala end moraine was used to reconstruct the past water level of the White Sea during the Younger Dryas Stadial. This procedure was partly based on Mann et al. (1999) description of reconstruction using the geographical information system (GIS) software application.

Relative water level history can also be reconstructed using conventional shore displacement methods (cf. Miettinen 2002), as presented in Paper II. These methods are based on identifying and dating the isolation contact from lake sediment cores obtained from small raised lake basins. The isolation contact, which can often been seen as a marked change in sediment type and diatom content, indicates the event when a lake basin becomes isolated from the ancient sea basin or a large freshwater body. The isolation contacts were dated here with a ^{14}C AMS dating method, using organic material collected directly above the isolation contact. The data obtained were used in Paper II for determining the threshold altitude and the time when the water level dropped below the threshold altitude and to reconstruct a shore displacement for the Belomorsk area.

The basic framework of the Fennoscandian deglaciation chronology was established using glaciolacustrine varve counting (De Geer 1912, Sauramo 1918, 1923) and biostratigraphical methods, combined with ^{14}C dating (cf. Hyvärinen 1973). During the past decade, methods involving luminescence (Lian 2007) and sediment exposure dating (cf. Ivy-Ochs & Kober 2007) have been also used to date the Younger Dryas Stadial end moraines (cf. Svendsen et al. 2004, Rinterknecht et. al. 2004, 2006). However, in Paper IV the deglaciation chronology for Russian Karelia was determined using varve counting together with palaeomagnetic dating (cf. Thompson & Oldfield 1986). The palaeomagnetic method is based on measuring the direction and intensity of Natural Remanent Magnetisation (NRM). Palaeomagnetic records of the studied lake sediments were compared against the Nautajärvi palaeomagnetic master curve (Ojala & Saarinen 2002) and the Mustalampi reference curve (Kosonen & Ojala 2011). The whole procedure, including echo sounding, coring, logging, sub-sampling and measuring of the palaeomagnetic samples and X-ray radiographs, is described in Paper IV. In the area east of the Kalevala and Pääjärvi end moraines, ^{14}C AMS dating, geomorphology and assumptions of ice lobes retreat rates were used to picture different deglaciation events, as outlined in Papers I and IV.

4 RESULTS: REVIEW OF PAPERS I–IV

4.1 Paper I

In Paper I, glacial geomorphological landforms in an area covering more than 7000 km2 were studied in detail using aerial photography and satellite imagery and on-site field observations. Drumlins, drumlinoids and megaflutes, non-streamlined landforms such as eskers and ice marginal landforms were manually drawn on 1:50 000 and 1:200 000 scale topographical maps. This was done to reconstruct the detailed history of Scandinavian Ice Sheet (SIS) behaviour and also deglaciation chronology in the Lake Kuittijärvi area. The SIS deglaciation chronology presented in Paper I is based on relative geomorphological data, ^{14}C AMS dating and assumptions on the ice retreat rate.

Based on the geomorphological data, three different ice flow patterns (two ice marginal zones and an interlobate complex) were identified in the area. The results indicate that the Tuoppajärvi sub-ice lobe (TIS) formed after the White Sea ice lobe became inactive. The Kuittijärvi sub-ice lobe (KIS), which was part of the Northern Karelian ice lobe, operated in the area during deglaciation. Subglacially formed lineation patterns associated with other indicative landforms, such as end moraines and esker ridges, indicate a clear age relationship between the activity of the ice lobes. The KIS was active after the linear landforms were formed by the TIS. It is estimated that deglaciation of the TIS from the Kalevala end moraine to the Lake Pääjärvi end moraine took place 11 300–10 900 calendar years ago. It seems that the ice margin of the KIS marked by the Kalevala end moraine was formed around 11 300–11 200 calendar years ago. During this time an interlobate area and later a passive ice area were formed between the active KIS and the retreating TIS. During the late deglaciation phase, the Kuusamo-White Sea ice lobe activated (the Pääjärvi flow stage) after the TIS became inactive and the Lake Pääjärvi end moraine was deposited ca. 10 900 calendar years ago. All three ice lobes (KIS, TIS and Kuusamo-White Sea) terminated in a relatively large ice lake extending from the Kalevala end moraine zone to the White Sea Basin. The level of this ice lake in the Kalevala end moraine zone and the Kepa area was approximately 125 m a.s.l. and in the Pääjärvi end moraine area approximately 145 m a.s.l.

4.2 Paper II

Paper II presents the results of the shore displacement study carried out in the Belomorsk area, NW Russia. A total of five small lake basins at altitudes of 72, 93, 113, 121, 134 m a.s.l. and one paludified pond at 45 m a.s.l. were cored and their sediments studied using traditional shore displacement methods. The purpose of the study was to identify and date the isolation contact of the lake sediment cores obtained from raised small lake basins. Each isolation contact, which indicates the event when the lake basin became isolated from the main water body, was dated by applying a ^{14}C AMS method to organic material collected from around the isolation contacts.

In most of the basins studied, the isolation contact was a sharp distinction where laminated or massive clay and silt, often poor in diatoms, changed into organic-rich clay or gyttja abundant in diatoms. Diatom analysis suggested that both minerogenic clay and silt and organic-rich clay and gyttja were deposited in a freshwater environment. In some of the basins the minerogenic silt and clay sediments included diatom taxa, indicating that the basins were isolated from a large glacial lake. At the isolation boundary the occurrence of Fragillaria species was very characteristic and above the isolation boundary the diatom taxa were typical for small lake basins. ^{14}C AMS dating of the isolation contacts of the lake basins at between 134 and 72 m a.s.l. showed an uncalibrated ^{14}C age of between 10 190 ± 60 and 10 290 ± 50 yr BP. The calibrated results indicated that there was an extensive ice lake in the White Sea Basin prior to ca. 12 050 cal. yr BP and initially all lake basins studied were part of this ice lake. Based on ^{14}C AMS data, diatom content and lithostratigraphical observations, it was shown that the lake basins between 134 m to 72 m became isolated from the White Sea basin ice lake within less than 200 years, around 12 050 cal. yr BP. There were suggestions that at that time, the water level of a large glacial lake in the Belomorsk area, which was part of a substantial glacial lake in the White Sea Basin area, experienced a sudden water level drop of approximately 50–60 m. As a result of this, a huge amount of glacial meltwater was released, most probably through the neck of the White Sea into the Barents Sea. The resulting sudden water level drop in the glacial lake in the White Sea Basin must also have affected the behaviour of the eastern flank of the Scandinavian Ice Sheet and the local climate around 12 000 cal. yr BP.

4.3 Paper III

Paper III presents the first numerical reconstruction of the White Sea Basin water level during the late Younger Dryas Stadial. In the study, geomorphological, sedimentological and ground penetrating radar survey (GPR) methods were used to study glaciofluvial plains and shorelines at four sites in the Kalevala end moraine area.

Geomorphological results for flat glaciofluvial plains and their steep distal slopes, discontinuous melt water channels and kettle holes were used to define the highest shorelines of the glaciofluvial formations studied. In addition to the geomorphological data, shallow test pits on flat plains and GPR indicated that three of the glaciofluvial plains represent glaciofluvial Gilbert-type deltas, deposited to the contemporary water level during the late Younger Dryas Stadial. The distance diagram for the study area was reconstructed using the data collected on shoreline altitudes, combined with shore line altitude observations by Ekman & Iljin (1991) and Putkinen & Lunkka (2008). The shoreline gradient was estimated to be 0.42 m/km and this value was used for numerical reconstruction of the White Sea water level during the late Younger Dryas Stadial. This reconstruction revealed that the water body in the White Sea basin was more extensive than today, inundating the present onshore areas on the western side of the White Sea and the Arkhangelsk area to the east. The ice margin terminated in the White Sea, which was connected to the Barents Sea via the Gorlo Strait and separated from the Baltic drainage basin to the south.

4.4 Paper IV

Paper IV presents the Late Weichselian and Early Holocene deglaciation history of the Kaleva and Pääjärvi area, Russian Karelia, together with absolute age determinations of the Kalevala and Pääjärvi end moraines. Determination of the deglaciation history of the area was based on the sedimentological and palaeomagnetic results obtained from three lake sediment cores. Two of the lakes, Ala-Kuittijärvi and Keski-Kuittijärvi, are situated on the proximal side of the Kalevala end moraine, while the third, Tuoppajärvi, is situated on the distal side of the Pääjärvi end moraine. One site from each lake basin was chosen for sedimentological and chronological studies. The chronology presented in Paper IV is based on palaeomagnetic measurements and varve counting results and comparison of these against the Finnish palaeomagnetic master curve.

The results indicate that the deglaciation sediments in Lakes Ala-Kuittijärvi and Keski-Kuittijärvi were deposited mainly by extra-marginal rivers, while those in Lake Tuoppajärvi were deposited in an ice-contact setting. After the glacial meltwater input ceased, typical large-lake gyttja clay/clay gyttja sediment accumulated in all three basins. The longest and the most complete chronology was obtained from Lake Ala-Kuittijärvi, for which the palaeomagnetic record extends back to 10 800 cal. BP. The sediment sequence from Lake Keski-Kuittijärvi covers at least 9 800 years and that from Lake Tuoppajärvi 9 800 years. Palaeomagnetic records together with 450 counted varves in Lake Ala-Kuittijärvi indicate that the basin was deglaciated 11 250 cal. BP and the Kalevala end moraine was formed 11 400–11 300 cal. BP while the Pääjärvi end moraine was formed 11 000 cal. BP. It was shown that the Kalevala end moraine is not correlated in time to the Younger Dryas Salpausselkä I and II moraines. It was suggested that the Kalevala end moraine was formed at the earliest Holocene at the same time as the Salpausselkä III and Pielisjärvi end moraines.

5 DISCUSSION

The Younger Dryas Stadial end moraines are important landforms for reconstruction of the deglaciation history of Fennoscandia and NW Russia. Glacial geomorphological features in Russian Karelia have been described in numerous maps and papers during past decades (cf. Kurimo 1982, Niemelä et al. 1993, Punkari 1985, Ekman & Iljin 1991, Rainio et al. 1995, Kleman et al. 1997, Boulton et al. 2001, Putkinen & Lunkka 2008). The most prominent end moraines in the western part of Russian Karelia are the Rukajärvi, Kalevala and Pääjärvi end moraines (see Fig. 1) (cf. Punkari 1985). It is thought that the Kalevala and Rukajärvi end moraines were deposited in front of the SIS during the Younger Dryas Stadial, ca. 12 800–11 500 years ago (Muscheler et al. 2008), although absolute age determinations are lacking.

Two different ice lobes, the Kuusamo-White Sea and Northern Karelian ice lobes, operated time-transgressively in the study area during the last deglaciation (Fig. 2) (cf. Punkari 1985, Ekman & Iljin 1991, Boulton et al. 2001). The ice margin of the Kuusamo-White Sea ice lobe was situated further east of the study area, i.e. in the White Sea Basin, when the streamlined landforms parallel to ice flow were formed in the Lake Tuoppajärvi area. Punkari (1985) concluded that the southern margin of the Kuusamo-White Sea ice lobe (i.e. the southern margin of the so-called Lake Onega sub-ice lobe) is marked by the Rukajärvi end moraine running from Rukajärvi to Belomorsk. Based on one OSL date, Lunkka et al. (in prep.) suggest that the eastern part of the Rukajärvi end moraine was deposited around 12 500 years ago (Fig. 4A). At this time, the Lake Onega sub-ice lobe terminated in a glaciolacustrine lake that occupied the Onega Bay of the present White Sea Basin (cf. Ekman & Iljin 1991, Rainio 1995, Yevzerov 1998) and lowland areas onshore to the south of the eastern arc of the Rukajärvi end moraine.

However, based on geomorphological observations which indicate that a part of the eastern arc of the Rukajärvi end moraine was deposited in the supra-aquatic setting (Punkari 1985, Ekman & Iljin 1991, Niemelä et al. 1993), it is evident that there were no major glaciolacustrine basins into which the ice front terminated in this area.

The beach deposits on the flanks of the Rukajärvi eastern arc end moraine ridges and the Rukavaara hill clearly indicate that there must have been transgression in the area when ice retreated towards the north. The transgression event that occurred in the glaciolacustrine basin post-dates the formation of the eastern arc of the Rukajärvi end moraine. Since there are no thresholds between the Rukajärvi hill and the White Sea Basin, which was already ice-free at that time, a major transgression took place in the White Sea basin and the adjacent onshore areas. On the Rukavaara hill the highest beach deposits occur at 175 m a.s.l., indicating the highest level of this ice lake (Fig. 4B). The results presented in this thesis suggest that the threshold for this glacial lake was situated in the present Gorlo Strait area to the northeast and in the Maaselkä upland area to the south (Fig 2).

As the ice retreated further from the Rukajärvi end moraine towards the north, the ice lake became larger and transgression continued until the new outflow route opened ca. 12 050 years ago, leading to an approximately 50–60 m regression in the glaciolacustrine basin within 200 years, as discussed in Paper II. The results suggest that the drop in water level in the glaciolacustrine basin had a major impact on the ice or sediment/sedimentary rock threshold in the Gorlo Strait area. This was eventually eroded as the waters of the White Sea Basin ice lake broke catastrophically through the threshold. Deglaciation continued as the ice margin of the former Lake Onega sub-ice lobe retreated to the northwest.

Geomorphological evidence indicates that the Northern Karelian ice lobe advanced subsequently from the west (cf. Punkari et al. 1985, Ekman & Iljin 1991, Niemelä et al. 1993). The direction of eskers and end moraines west of Tiksha clearly shows that the Northern Karelian ice lobe overrode the westernmost part of the eastern Rukajärvi end moraine arc (Fig. 4C) (cf. Niemelä et al. 1993, Putkinen & Lunkka, unpublished). It has also been suggested that the western arc of the Rukajärvi end moraine was deposited during this ice advance (cf. Punkari 1985, Rainio et al. 1995, Lunkka et al. 2004). It seems evident that the northern part of this end moraine in the Tiksha area was deposited supra-aquatically (Putkinen & Lunkka, unpublished). Since the end moraines in the Tiksha area are lower than the highest shore line altitude of the adjacent sub-aquatic Rukavaara hill area, the deposition of the end moraines in the former area occurred after the water level drop in the White Sea basin Ice lake ca. 12 050 years ago.

As the ice margin retreated to the Kalevala area, the geomorphological evidence, particularly the fact that longitudinal eskers run from both the Tuoppajärvi sub-ice lobe and the Kuittijärvi sub-

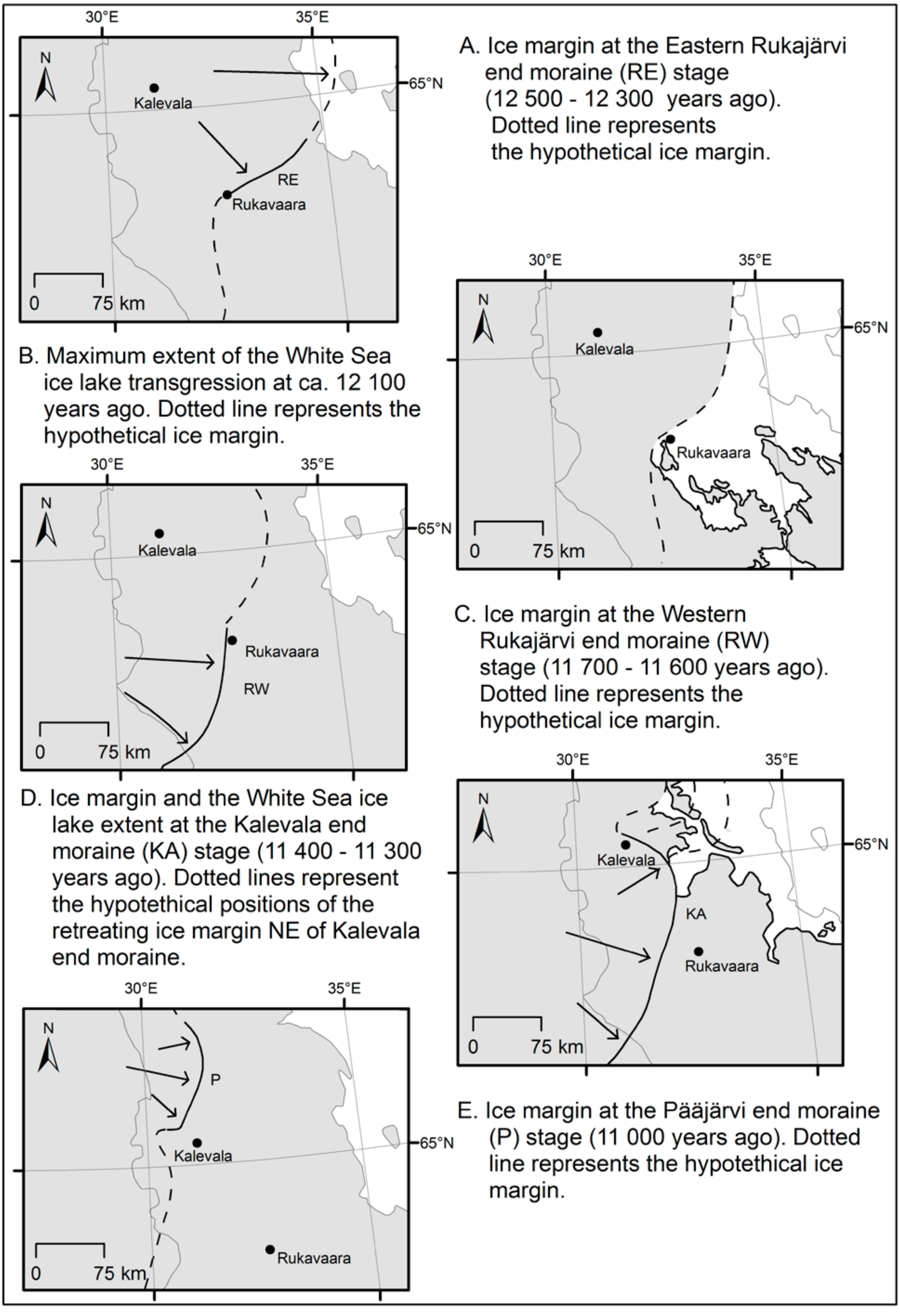

Fig 4. Late Weichselian and Early Holocene deglaciation of Russian Karelia. The palaeoshoreline of the ice lake that existed in the White Sea Basin is also presented in figures B and D.

ice lobe areas, indicates that interlobate formations were deposited between these two ice lobes (Fig. 2). The positions of the ice margin in the area of the Kuittijärvi sub-ice lobe and the Tuoppajärvi sub-ice lobe can be reconstructed from the deltas that formed in front of the ice margin in the glacial lake (Figs. 2 and 4D). The glaciofluvial deltas with feeding eskers south of Shonga and in Kepa lie at approximately 125 m a.s.l. and both areas are approximately at the same south-north trend of the modern isobase. This means that these two areas have experienced the same uplift history after the Weichselian deglaciation. It is therefore most likely that the deltas south of Shonga and around Kepa were deposited in the glacial lake that existed next to the ice margin. This ice lake must have been part of a large ice lake that extended to the present White Sea basin (Paper II).

As the geomorphological evidence suggests, the Tuoppajärvi sub-ice lobe retreated before the Kuittijärvi sub-ice lobe as deglaciation proceeded (Fig. 4D). Based on the glaciofluvial delta levels between Kepa Village and Lake Suolajärvi in the area of the Tuoppajärvi sub-ice lobe, it seems that the ice margin terminated in this large ice lake, the highest shoreline of which was at 145 m a.s.l. in the area of Lake Suolajärvi. Delta levels higher than 125 m a.s.l. are not present in the area of the Kuittijärvi sub-ice lobe, which together with ^{14}C AMS dating results suggest that it still occupied the low land area of the Lake Kuittijärvi basin and its terminus was at the present Kalevala end moraine and continued along the interlobate formation to the north and west (Fig. 4D). Longitudinal eskers, deltas, sandur deltas, sandar and the kame-kettle topography between Kepa and Lake Ohdanjärvi were deposited during the course of the time-transgressive retreat of the Tuoppajärvi sub-ice lobe. Thus these landforms were not formed in the collision zone of the above ice lobes, as previously suggested by Punkari (1985).

It also seems evident that the Kuusamo-White Sea ice stream, including the Tuoppajärvi sub-ice lobe, did not operate with a surge fan mechanism, as proposed by Kleman et al. (1997). This conclusion is supported by the fact that there are end moraines and eskers feeding into the ice marginal deltas in the onshore area between the present White Sea Basin and the Pääjärvi end moraine, without any reorganisation of the streamlined, subglacial features. This clearly indicates a successive and time-transgressive retreat of the ice margin. The Lake Pääjärvi drumlin field and associated end moraine were formed during the Lake Pääjärvi flow stage of the Kuusamo-White Sea ice stream (Fig. 4E). This flow stage was considerably younger than the Tuoppajärvi flow stage that formed the Tuoppajärvi drumlin field.

The shore displacement study in the Belomorsk area, together with the geomorphological evidence from the Tiksha area, show that the ice margin followed the Rukajärvi end moraine eastern arc from Tiksha to Belomorsk. This end moraine was deposited prior to the transgression of the White Sea ice lake. This means that the Rukajärvi end moraine eastern arc was deposited slightly earlier than the 12 300–12 100 years ago suggested by Saarnisto & Saarinen (2001), possibly 12 500–12 300 years ago. According to the geomorphological data discussed above, the western arc of the Rukajärvi end moraine from the Tiksha area to the Koitere end moraine in Finland was deposited after the formation of the eastern arc of the Rukajärvi end moraine. The western arc of the Rukajärvi end moraine was laid down in the time interval between the rapid ice lake regression 12 050 years ago and the formation of the Kalevala end moraine 11 400–11 300 years ago. Assuming an ice retreat rate of approximately 300–400 m/year during the last deglaciation (Boulton et al. 2001), this end moraine was formed at least 100–150 years before the Kalevala end moraine. Based on this, it is more likely that its age is 11 700–11 600 years, corresponding roughly to the age of Salpausselkä II and the Koitere end moraine in Finland.

In the Kalevala end moraine area, 150 km north of Tiksha, the palaeomagnetic dating and varve-counting results from the Ala-Kuittijärvi and Keski-Kuittijärvi lake sediment sequences indicate that the ice had already retreated from the Kuittijärvi basin 11 250 years ago. Since the western margin of Ala-Kuittijärvi is situated approximately 30 km inside the Kalevala end moraine, the end moraine itself must be older than 11 250 years. Assuming that the ice retreat rate in northwestern Russian Karelia was approximately 300–400 m/year during deglaciation, as suggested by Boulton et al. (2001) for example, then the age estimate for the formation of the Kalevala end moraine would be ca. 11 400–11 300 years ago.

Recent age determinations of the Salpausselkä I and II end moraines in Finland, adjacent to the Russian Karelian Kalevala and Pääjärvi end moraines discussed here, indicate that the Salpausselkäs were deposited during the latter part of Younger Dryas Stadial and the earliest Holocene, 12 250–11 590 years ago (cf. Saarnisto & Saarinen 2001, Tschudi et al. 2000, Rinterknecht et al. 2004). According to the present results, the Kalevala end moraine was formed somewhat later than the Salpausselkä II end moraine (Fig. 1). Previously, Andersen & Borns (1997), Rainio et al. (1995) among others, correlated the Kalevala end moraine with the Pielisjärvi end moraine in Finnish northern Karelia. This correlation scheme is supported by the results presented here. It is also suggested that the Salpausselkä III in SW Finland was formed at the same time, i.e. 11 400–11 300 years ago.

Palaeomagnetic results from Tuoppajärvi, situated immediately on the distal side of the Pääjärvi end moraine, extend only 9 500 years back in time. Therefore, this age is only the minimum age for the formation of the Pääjärvi end moraine. Based on the ^{14}C AMS dates shown in Paper I, the area 10 km on the distal side of the Pääjärvi end moraine had already become ice free ca. 10 800 years ago and the authors conclusion was that the Pääjärvi end moraine was formed ca. 10 900 years ago. Due to the fact that the Pääjärvi end moraine is situated around 80 km northwest of the Kalevala end moraine measured along the interlobate system and that it was formed at the margin of a different ice lobe than the Kalevala end moraine (Fig. 1), the Pääjärvi end moraine must have formed later than the Kalevala end moraine. The age estimate of 11 000 years ago presented here is one hundred years older than previously estimated in Paper I. This refinement is due to more precise absolute dating results of the Kalevala end moraine together with the assumption that the ice-retreat rate was 300–400 m/year over a distance of 80 km.

6 CONCLUSIONS

Deglaciation of Russian Karelia took place time-transgressively as the SIS retreated from the area during the Younger Dryas Stadial and the earliest Holocene. A great variety of glacial landforms were formed from which the deglaciation history of the area can be reconstructed. The main conclusions presented here are based on mapping of these glacial landforms using aerial photographs, satellite imagery and field investigations. The chronostratigraphy and geochronology are based on geomorphological considerations, ^{14}C dating and palaeomagnetic dating of annually laminated large lake sediments cored from the end moraine zones. A shore displacement study was conducted in the Belomorsk area, together with modelling of the water body of the ice lake that existed in White Sea Basin to gain an overall picture of the events and environmental changes that took place in Russian Karelia during the Late Weichselian and Early Holocene periods. Based on the evidence presented in this thesis, the following conclusions can be drawn:

1. The Rukajärvi end moraine eastern arc was deposited at the ice margin of the Onega sub-ice lobe 12 500–12 300 years ago.
2. A large ice lake then began to form 12 300 years ago and occupied the White Sea Basin and onshore areas. The transgressive phase of this ice lake ended 12 050 ± 150 years ago, when its water level dropped rapidly by around 50–60 m due to the opening of a new drainage route via the Gorlo Strait.
3. The Rukajärvi end moraine western arc was deposited at the margin of the Northern Karelian ice lobe 11 700–11 600 years ago.
4. As deglaciation proceeded, the configuration of the ice lobes changed, which led to collision between the active Kuittijärvi and more passive Tuoppajärvi sub-ice lobes and the deposition of the Kalevala end moraine 11 400–11 300 years ago.
5. After the interlobate deposition of the northern part of the Kalevala end moraine, ice retreated from the north of the interlobate zone more rapidly than for the Kuittijärvi sub-ice lobe to the south. After the retreat of the Kuittijärvi sub-ice lobe from the Kalevala end moraine, the Pääjärvi end moraine was formed at the margin of the Kuusamo-White Sea ice lobe 11 000 years ago.

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This thesis comprises a synopsis and four original papers dealing with the deglaciation history and chronology of the eastern flank of the Scandinavian Ice Sheet, and the evolution of the White Sea Basin in NW Russia during the Late Weichselian. The studies included mapping of the geomorphological formations and determination of the age relationship between different landform associations. This information was used in the reconstruction of the dynamics of ice lobes in the area. In addition, studies on lake sediments and shore displacement, as well as sedimentological and geophysical investigations were carried out to reconstruct the palaeoenvironments that existed in the White Sea Basin in the Late Weichselian. The results indicate multistage evolution of the Scandinavian Ice Sheet in the study area, where the Kalevala and Pääjärvi end moraines, geomorphologically correlative to the Younger Dryas Stadial Salpausselkä end moraines, were formed shortly prior to 11 400–11 300 and 11 000 years ago, respectively. The age estimates presented in this thesis suggest that the Kalevala and Pääjärvi end moraines were deposited 100 to 500 years earlier than previously thought, i.e. their deposition post-dates the deposition of the Second Salpausselkä end moraine in SE Finland. It is also shown that an extensive glacial lake existed in the White Sea Basin during the early stages of deglaciation. This glacial lake became connected to the Barents Sea at 12 050 years ago, when the water level of the glacial lake rapidly dropped by ca. 60 m. It is suggested that a huge amount of fresh water entering the Barents Sea at 12 050 years ago, together with the freshwater input from the ca. 28 m drop in the level of the Baltic Ice Lake slightly later, must have had an impact on the climate all around Scandinavia and NW Russia.

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