The Last Scandinavian Ice Sheet in northern Russia: ice flow patterns and decay dynamics

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Invasion of the Late Weichselian (Valdaian) Scandinavian Ice Sheet (SIS) in northern Russia took place after a period of periglacial conditions. Till of the last SIS – Bobrovo till, overlies glacial deposits from the previous Barents and Kara Sea ice sheets and marine deposits of the last interglacial. The till is identified by its contents of Scandinavian erratics and it has directional properties of westerly provenance. Above the deglaciation sediments and extra marginally it is replaced by glaciofluvial and glaciolacustrine deposits.

At its maximum extent, the last SIS was more restricted in Russia than previously outlined and the time of termination at 18-16 cal kyr was almost 10 kyr delayed compared to the southwestern part of the ice sheet. We argue that the lithology of the ice sheets substrate and especially the location of former proglacial lake basins influenced the dynamics of the ice sheet and guided the direction of flow. We advocate that while reaching the maximum extent, lobe shaped glaciers protruded eastward from SIS and moved along the path of water filled lowland basins. Collapse and deglaciation in the region commenced when ice lobes were detached from the main ice sheet. During the Late Glacial warming, disintegration and melting took place in a 200 – 600 km wide zone along the north-eastern rim of SIS associated with thick Quaternary accumulations. Deglaciation occurred through aerial downwasting within large fields of dead ice developed during successively detached ice lobes. Deglaciation led to the development of hummocky moraine landscapes with scattered periglacial and ice-dammed lakes, while a sub-arctic flora invaded the region.

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The eastern flank of the Scandinavian Ice Sheet (SIS) occupied larger parts of the Arkhangelsk district of the Russian European North during the Late Weichselian (Valdaian). Although traces left by the SIS have been recognized since the turn of the last century (Ramsay 1911), the ice flow dynamics and distribution of landforms and sediments generated near the eastern limit of the maximum extent of SIS in northern Russia is still at debate (Svendsen *et al*. 2004). The duration and age of the glacial maximum is significantly different compared with the ice sheet's termination nearer to the Atlantic continental margin. Thus, it appears that the peak glaciation in Russia was delayed by almost 10 kyr. In southwest Scandinavia this state was reached sometime between 32 – 25 kyr (Houmark-Nielsen & Kjær 2003; Sejrup *et al.* 1994, 2003) when SIS eventually merged with the British Isles Ice Sheet. On the northern boundary SIS coalesced with the Barents Sea Ice Sheet around 22 kyr (Svendsen *et al.* 2004, and references therein). The exact location of the merging is less evident largely due to lack of mappable landforms and the restricted stratigraphical control of offshore sediments. Adding to these uncertainties, the mode of advance and decay in time and space of the northeastern sector of the Scandinavian Ice Sheet has until recently been poorly documented (Demidov *et al*. 2004).

In this paper, a revised version of Late Weichselian ice sheet flow dynamics, deglaciation pattern and the development of landforms is presented from the time of maximum extent of the

Scandinavian Ice Sheet in NW Russia until the beginning of the Holocene. This was achieved through a critical analysis of previous investigations of the Quaternary stratigraphy and geomorphology supported by new field observations and new age constraints. Data from the northern Arkhangelsk Region (Fig. 1) including the Kanin Peninsula, the seashore of the Kuloi Plateau and bluffs along the Pinega River are synthesized with previously published data from the Severnaya Dvina and Mezen rivers (Larsen *et al*. 1999; Lyså *et al*. 2001; Kjær *et al*. 2003) (Fig. 1). From 1996 to 2002 more than 100 key sections with Late Pleistocene deposits were investigated by stratigraphical, sedimentological and palaeontological methods along well exposed cliff sites on the seashore and riverbanks in the Arkhangelsk Region. By addressing these issues we close a gap in the knowledge about Late Weichselian glaciation history along the fringes of the whole of the last SIS.

Setting

The study area is located on the southern and eastern shores of the White Sea, along the Kanin Peninsula and the Kuloi Plateau and it covers the middle part of the Pinega River and the lower part of the Mezen River (Figs 1, 2). According to Geology of USSR (1963) a rugged denudation plain, composed of Palaeozoic sediments are dissected by the 100-270 metres high water divides of the Kuloi Plateau and the Pokshenga Highland, whereas the Kanin Ridge is composed of Late Precambrian meta-sediments and small amounts of igneous rocks (Figs 1, 2). The Severnaya Dvina, Pinega, Mezen and Kuloi rivers and the root of the Kanin Peninsula occupy extensive depressions, with elevations from 10-80 metres above sea level. Deep troughs in the White Sea lying more than 50 metres below present sea level separate the Palaeozoic sedimentary platform from the Precambrian crystalline basement of the Fennoscandian Shield. Apart from more than 100 m deep incised valleys, the Quaternary cover is 5-60 metres thick except for uplifted regions, where strongly weathered bedrock and boulders fields are present (Fig. 2). Glacigenic deposits comprising till, fluvial and lacustrine sediments of the Late Weichselian Scandinavian glaciation dominate subsurface exposures in the western part of the discussed region, whereas the areas to the east were glaciated by the Barents and the Kara ice sheets in Early and Middle Weichselian time (Kjær *et al.* 2003; Larsen *et al.* 2006a).

A comprehensive review of previous investigations in northwestern Russia was provided by Demidov *et al*. (2004). Ramsay (1911) recognized three till units separated by shell bearing sediments on the Kanin Peninsula, at the mouth of Mezen River and on the Kuloi Plateau Seashore. Using the lithology of erratic boulders, Ramsay proposed that Scandinavia and Novaya Zemlya were centers of glaciation. Two or three till beds have since then been recognized in the Kanin Peninsula area (Kalyanov & Androsova 1933; Lyutkevich 1947; Spiridonov & Jakovleva 1961), the latter authors also supported the idea that the upper till was of Scandinavian origin. This is confirmed by Kjær *et al.* (2006) and Larsen *et al.* (2006b). The presence of Scandinavian boulders, young eskers and sandy kames in the western parts of Kanin Peninsula, led Devyatova (1969), Aseev (1974) and Demidov *et al*. (2004) to support the contention of Ramsay (1911) that the SIS reached the central axis of the Kanin Peninsula during the last glacial maximum (Fig. 1). From the shores of Mezen Bay, the boundary of the Late Weichselian SIS has been drawn southward towards Severnaya Dvina via the central part of the Kuloi Plateau (Krasnov 1971; Legkova & Schukin 1972; Ganeshin *et al.* 1980), although Lavrov (1991) slightly modified these reconstructions using aerial photopraphic interpretation. Edemsky (1931), Devyatova (1982), Filippov and Borodai (1987) suggested that the middle parts of the Pinega and Mezen rivers remained unglaciated after the Eemian (Mikulinian) interglacial, while others argued that SIS or Kara Sea Ice Sheet reached the area either during the Early Weichselian (Krasnov 1971) or at the end of the Weichselian (Lavrov 1991; Aseev 1974; Lavrov & Potapenko 2005).

Methodology

Our geomorphological interpretation is based on analyses of topographic and geological maps of scale 1:100 000 and 1:200 000, satellite images of scale 1:500 000 and 1:1000 000 (Zatulskaya 1984 a, b, c). Field observations in selected areas served as ground control. In Figure 2 we have distinguished between end moraines, hummocky moraine, kames, sandar, glacial lake basins, bed rock or weathered rock, bed rock cliff, valley trains and plateaus.

River and coastal sections were logged and facies associations were interpreted in terms of depositional environment. Diamictons were described, classified and interpreted following common guidelines (c.f. Krüger & Kjær 1999; Kjær *et al.* 2001). Directional properties of till units include striations on sub-till bedrock and boulder pavements and most frequently, data from clast fabric analyses. Samples of 25 clasts were obtained from a sub-horizontal surface *c.* 25x25 cm and only clasts with a long-to-intermediate axis ratio ≥1.5 and a length between 0.6 and 6 cm were measured. Eigenvectors (V_1) and eigenvalues (S_1 and S_3) for each sample were computed using the program SpheriStat, version 2.0 (Pangaea Scientific) based on Mark (1973, 1974). According to Anderson and Stephensen (1971) clasts are considered to have statistical preferred orientation if, respectively, the S1and S3 eigenvalues are ≥0.52 and ≤0.17. In order to verify the significance of eigenvectors and test for bimodal distribution of clasts, contoured diagrams were contoured following the approach by Kamb (1959).

When unit boundaries comprise angular unconformities indicating glaciotectonic deformation such as shearing, folding and thrusting, a deformational chronology including the direction of glacier movement is added to the stratigraphic record (Houmark-Nielsen & Kjær 2003). Fine gravel counts were carried out in the 3-8 mm fraction and split into different provenance dependent rock types, which enables discrimination of tills from glaciers that originated on the Fennoscandian Shield to the west from those originating on the Palaeozoic platform and Mesozoic uplifts to the east and northeast (Kjær *et al.* 2001).

Relative chronological control was obtained by using the Eemian marine sediments as a lower marker bed (Devyatova 1982; Larsen *et al.* 1999; Funder *et al.* 2002), and by distinguishing tills of Scandinavian origin from older Weichselian tills of Barents and Kara Sea origin by their content of provenance dependent erratics and other directional properties (Houmark-Nielsen *et al*. 2001; Kjær *et al*. 2001, 2003, 2006; Larsen *et al.* 2006a, b). Till of successive advances of SIS contain erratic clasts from the Fennoscandian Shield such as Precambrian granites, gneisses, basic rocks and quartzite from the Karelia-Kola province and especially Kola Peninsula nepheline syenites are found dispersed in the northern part of Arkhangelsk district. Reddish, Late Proterozoic sandstone from the southern seashore of the Kola Peninsula also occurs in till deposited by the Scandinavian Ice Sheet and in the fine gravel fraction such till is almost devoid of limestone and dominated by crystalline rock fragments followed by Lower Palaeozoic sandstone, mudstone and shale.

Luminescence dating was carried out at the Nordic Laboratory for Luminescence Dating, Risø, Denmark. The results are listed in Table 1. Dose measurements were made using quartz and a Single Aliquot Regenerated dose protocol (Murray & Wintle 2000). Dose rate analysis is based on high resolution gamma spectrometry (Murray *et al*. 1987). Calculations used a saturation water content of 25 - 30%. These are typical for such sandy sediments, but two samples had measured water contents of 6 - 7 % lower than this. Using these values would only reduce the ages by 5 - 6%, and since these lower saturated values are considered unlikely, we have chosen to use the higher range for all samples, with an assigned uncertainty of ± 4%. It is assumed that the sediments were saturated for more than 95% of the time after burial (Larsen *et al*. 1999; Houmark-Nielsen *et al*. 2001; Kjær *et al*. 2003). We base this assumption on the deduction that the sediments were frozen during most of the Weichselian; in periods of thaw they would have remained in the saturated ground water zone – until sub-recent erosion by modern rivers dropped the local ground water table. The sample burial depth also affects the age, because of attenuation of cosmic rays by overburden sediments. The lifetime averaged burial depth is, of course, not known. Although the present depths beneath the surface not necessarily equal the averaged burial depths over time we have assumed a constant burial depth of 3.9 m. Had we used observed depths, some of which can be deduced from Figures 3 and 7, the ages would only have changed, on average, by 0.3% (s=4%). This mean change is very small compared to the typical overall age uncertainty of ~10%, as is the standard deviation in this change. All uncertainties resulting from these possible variations in burial depths are <55% of the total uncertainty in the ages, and those ages critical to the dating of the last glacial maximum (those between 30 and 12 ka) are changed by <1000 years. In summary, the likely uncertainties arising from variations in water contents and sample depths are considered unlikely to add significantly to the overall uncertainties given in Table 1.

Radiocarbon dating of plant macro fossils was done at the AMS-facility at the Ångström Laboratory, University of Uppsala, Sweden and the Laboratory for Radiometric Dating, University of Trondheim, Norway (Larsen *et al.* 1999). The ^{14}C age is calibrated to calendar years after the calculations of Kitagawa & van der Plicht (1998). Rock samples for *in situ* cosmogenic ^{10}Be surface exposure dating (CED) were collected and treated after the procedures described in Linge *et al.* (2006). Ages are expressed in thousands of years (kyr) and indicated in stratigraphical logs and listed in Figure 9 and Table 1.

Maximum extent and age of the last Scandinavian Ice Sheet

The last SIS covered the northern coastal areas of the Kuloi Plateau, and reached the northwestern part of the Kanin Peninsula, and the central part of the Pinega River (Figs 1, 2). From these areas, stratigraphical logs and cross sections containing information on glacial deposits with Scandinavian provenance are compiled and used as a prominent marker. Detailed knowledge on the last SIS was previously presented from the Severnaya Dvina area (Larsen *et al.* 1999; Kjær *et al.* 2001; Lyså *et al.* 2001) where till of that ice sheet was named Bobrovo till. The Volodga area to the south of the Arkhangelsk Region was accounted for by Lunkka *et al.* (2001) and areas to the east that were not overridden by the ice sheet are described by Houmark-Nielsen *et al*. (2001).

Northern coast of the Kuloi Plateau – Eight sections from the northwestern corner of the Kuloi Plateau to the Mezen Bay coast at the root of the Kanin Peninsula comprise a stratigraphical cross section through a segment of the marginal zone of the last SIS (Figs 1, 3). Marine Eemian sediments underlie Weichselian glacial deposits in the river valleys draining the northwestern part of the Kuloi Plateau (Geology of the USSR 1963) just as sections with similar setting are known from the opposite shore on the Kola Peninsula (Armand *et al.* 1969). Previous detailed stratigraphic studies from the Mezen Bay are adopted from Kjær *et al*. (2003).

Till of former Weichselian glaciations and fluvial deposits underlie glacial deposits of Scandinavian origin. Ages range between 77 kyr and 21 kyr (Table 1). Scandinavian till unconformably overlies these sediments and may have been glaciotectonic deformed and the basal part of the till often shows sheared and folded lenses of Quaternary deposits. Sub-till and inter-till glaciodynamic structures, striations on clasts and clast fabric orientation indicate glacier movements from westerly and southwesterly directions (Fig. 4; sites 8, 9, 11). Occasionally, ice dammed lake sediments of similar stratigraphic position contain drop stones of Scandinavian origin (Fig. 3; sites 8-13, 15, 49). The till ranges in thickness from 6 to 2 metres and the clast content is characterized by a mixture of local bedrock and rock fragments of westerly provenance.

Glaciofluvial sand and gravel beds overlie the Scandianvian till or they constitute its lateral equivalent. Deposits also show ice wedge casts and other periglacial structures overlain by fluvial deposits containing plant detritus. Ages of these sediments yield 13 and 11 kyr (Table 1). Often the examined sections are covered by Holocene peat. Because the presence of Scandinavian till can be demonstrated in sections found at Cape Kargovsky and westwards, we assume the limit of SIS to be located near the east bank of the mouth the Kuloi River (Figs 1, 3). Even though the topography is rather flat and glacial landforms are blurred by peat bogs, an eight km long chain of end-moraines and hills that cross the Kuloi River about 30 km southwest of the Cape favours our proposed maximum position (Fig. 2).

Western coast of the Kanin Peninsula – Demidov *et al.* (2004) supported earlier views on the limit of SIS based on the distribution of Scandinavian erratics, young eskers and hummocky moraines by Ramsay (1911), Devyatova, (1969) and Aseev (1974) as indicated in Figures 1 and 2. However, our present understanding suggests the distribution of the Scandinavian till to be restricted only to the section near Tarkhanov southeast of the Cape Kanin Nos. This is under the assumption, that OSL dated glaciofluvial and glaciolacustrine sediments with ages older than 50 kyr have been well bleached (Fig. 1 sites 4-7, Table 1). These deposits overlie a till plain from earlier Weichselian glaciations at Konushin and other coastal sections of the western and northern Kanin Peninsula gave.

At the Tarkhanov cliff section (Fig. 1, site 1) a 40-45 metre high terrace of Weichselian sediments is banked upon the Precambrian meta-sediments that are inclined up to 90° (Fig. 5). On top of the exposure the Kanin Ridge is more or less flat and almost without Quaternary deposits. Strongly weathered bedrock, tors and fields of residual boulders are widespread, but occasionally 10-15 metre high kame-like hills composed of bedded silt and sand occur. In the beach terrace marine Eemian sediments are overlain by glacial and glaciofluvial deposits from the Early and Middle Weichselian dated to 87 kyr, 80 kyr and 58 kyr (Table 1). Boulders of crystalline Scandinavian rocks occur both on the present-day beach and on the deflation surface of the 35 metres high terrace. A brown sandy till unit less than 2 m thick is preserved in patches below the deflation surface (Fig. 6). Clast fabric orientations indicate ice-movement from the NW (Fig. 4). The till is rich in Scandinavian rock fragments and has a sharp erosional contact to the underlying lacustrine sand.

According to the lithology of clasts this till unit is equivalent to the Bobrovo till deposited by the last SIS further south. Cosmogenic nuclide dating of quartz veins on the bedrock surface at different altitudes indicate ages of exposure at 18.6 kyr below *c.* 50 metres a.s.l., whereas at adjacent inland altitudes around 100-115 metres a.s.l. ages indicate exposure around 51 kyr and 56 kyr (Linge *et al.* 2006). OSL dates of sediments in the kame hills at altitudes above 130 metres on the inland plateau a few km northward gave OSL ages of 71 kyr and 63 kyr (Table 1, Fig. 5). Aerial photographs show a ridge winding the NW-tip of the Kanin Peninsula at altitudes about 40-50 metres a.s.l along an escarpment in the bedrock. We interpret this ridge to be a type of terminal moraine associated with the latest glacier cover over Kanin Peninsula.

Thus, the upper till in the Tarkhanov section was most probably deposited by Scandinavian ice sheet the flow direction of which was either defflected by the advancing Barents Ice sheet from the north or the direction reflects local advance-phase flow influenced by the steep slope bedrock topography. Our data suggests that the last SIS reached an altitude up to 60-70 metres a.s.l. and moved along the White Sea slope of the Kanin Ridge before 18 kyr. The top of the Kanin Ridge, which reaches altitudes of 100-180 metres a.s.l., and because older sediments lack till cover at sites 2-7, we assume that this part of the Kanin Peninsula was ice free during the Late Weichselian.

Central Pinega River – Demidov *et al*. (2004) proposed that the Pinega ice lobe covered a significant part of the Pinega-Mezen watershed. New data from sections at Ezhuga, Karpogory and Verkola (Fig. 1; sites 19, 20, 22 and Fig. 7) suggests, however, the limit to be drawn about 50 km further to the west. The thickness of the Quaternary cover in the Pinega lowland is limited to less than 5-10 metres. Sediments with Eemian marine shells lacking a till cover are widespread in surface exposures (Fillipov & Borodai 1987). Near Karpogory, a section lying south of the pronounced end-moraine ridges crossing area from NE to SW (Figs 1, 2), a red-brownish clayey till with Scandinavian and local clasts is present. Fabric analyses in the till (Fig. 4) and age of the glaciofluvial and glaciolacustrine sediments below and above (Fig. 7) suggests the till to have been deposited by an ice sheet from north-northwest in the Early Weichselian or possibly in the Late Saalian. On the left side of Ezhuga River, marine interglacial sediments rest on the Moscowian (Late Saalian) Scandinavian till and is covered by re-deposited diamicton containing both local and Scandinavian boulders (Fig. 7).

Devyatova (1969, 1982) suggests the limit of the Late Weichselian SIS to have been located near the Karpogory village and we propose that the limit possibly is marked by the many end-moraine ridges and meltwater channels that occupy the north-western part of the area (Zatulskaya 1984a, b). Glaciolacustrine sand at Ezhuga (site 19) with an age of 26.6±1.9 kyr that lies above clast bearing mud (Fig. 7, Table 1) does not give any conclusive evidence as to weather the last SIS did override this area. West of the end-moraine belt i.e. east of the Kulogora village (Fig.1; site 18) till of Scandinavian provenance rests on silt and sand with an age of 77±7 kyr containing marine shells (Fig 7, Table 1). The till is covered by aeolian sand dated to 15±1.6 kyr. Clast fabric indicates ice movement from the west-southwest (Fig. 4). Our data from the central part of the Pinega River suggests the limit of the last SIS in this area to be corrected (Figs 1, 2, 7), however, the stratigraphic significance of the Scandinavian erratics in this area is limited. Till with a Scandinavian clast signature is present in most of the area, and such diamicts are situated both above and below sandy beds of redeposited marine shells of Eemian age.

The position of the last SIS in the Arkhangelsk region – Since Scandinavian till apparently is absent in sections along the coastal area of western Kanin Peninsula, the ice margin continued from the Cape Kanin Nos probably across the eastern White Sea floor southward to the Cape Kargovsky (Figs 1, 2). The Cape Kargovsky is the eastern most point in the Mezen Bay area, where Bobrovo till is observed (Kjær *et al.* 2001, 2003). From this point, the ice margin ran southward through the waterdivide between the Kuloi and Mezen Rivers and crossed the valley of the Kuloi River near the mouth of the Soyana River. The ice front apparently turned westward from the mouth of the Soyana River and touched the steep, 50-80 metre high, northeastern cliffs of the Kuloi Plateau, the northeastern part of which is deprived of Quaternary deposits. Discontinuous belts of sandy end-moraine hills constitute the western boundary of outwash plains, which guided melt-water towards the east (Krasnov 1971, Legkova & Schukin 1972).

During the maximum extent, the 180 metres high south-eastern part of the Kuloi Plateau was ice free (Edemsky, 1931), and we propose that the ice margin only rimmed the plateau. From the south an ice tongue moved eastward along the Pinega River lowland up to altitudes of 30-60 metres a.s.l., and the ice front covered only the middle part of the Pinega River (Figs 1, 8). The ice front ran along the terminal moraine belt. It expanded to the south and crossed the Pinega River and found its limit north of the villages Shilega and Karpogory, where Mikulinian marine sediments with whale bones are uncovered by till (Devyatova 1982). The ice margin continued along a discontinues chain of hills and ridges southwestward to join terminal formations, which truncate the Pokshenga Hills from the north and west between site 21 and 24 (Figs. 1, 2) (Arslanov *et al.* 1984). The maximum limit is connected with the limit of

SIS in the Severnaya Dvina, which is drawn for similar reasons by Larsen *et al.* (1999) and Atlasov *et al.* (1978).

In our present configuration of the last SIS, data suggests a reduced distribution of glacier ice and consequently its eastern boundary is moved westward by 20-120 km compared to previous reconstructions. We recognize that the western coast of the Kanin Peninsula, the River Kuloi and the eastern part of the Kuloi Plateau and areas situated eastward and southward from the Karpogory village in middle Pinega River were ice free during the last glaciation, but may have been within the reach of previous Scandinavian Ice Sheets. Lowland areas were occupied by large partly ice dammed lakes (Figs. 8, or Fig.10A-D). An almost straight crested, 130-km long belt of hills can be traced from the mouth of the Pyoza River, southward through the Kuloi River and Mezen waterdivide. It was proposed to mark an end-moraine zone (Lavrov 1991, Zatulskaya 1984c) and according to Demidov *et al*. (2004) it positioned the maximum extent of SIS. However, at exposures close to the mouth of the Pyoza river we have not identified any end moraine ridges or Scandinavian till within this belt, on the contrary, this straight ridge is found merely reflect the strike of gently eastward dipping bedrock.

A fundamental assumption in our reconstruction is that ice sheets deposit till or leave a tectonic imprint that indicate ice transgression, hence the distribution of a till sheet or glaciotectonic deformations might be used to reconstruct the extent of the ice sheet. In our opinion, it seems likely that wet-based ice lobes extended eastwards during LGM build-up. However, frozen bed conditions could have been regained in the marginal zone, when ice lobes reached higher and permafrzosen ground east of the White Sea basin, as it seems to be the case with the ice cover on the high grounds of the Kola Peninsula (Hättestrand 2006). Although, we are not able to exclude this possibility, there is compiling evidence for an extensive proglacial meltwater drainage system associated with the suggested ice sheet configuration in northwest Russia.

Age of the maximum extent of the last SIS – OSL dates above and below the Bobrovo till in the Arkhangelsk region supplemented by dates from Lake Beloe in the Volodga Region from similar stratigraphic levels (Lunkka *et al*. 2001) provide an age frame for the maximum extent (Fig. 9). Sub-till dates range from 32 kyr to 15.9$\pm$1.2 kyr whereas dates from above the Bobrovo till range from 17.2$\pm$1.3 kyr to 11 kyr. Allowing the ice sheet to settle at the maximum position for at least 1 kyr, the assumed age of this event lies between 18-16 kyr $\pm$ 1 kyr. Sediments outside the former ice margin indicate that damming of substantial lowland areas by ice marginal lakes commenced about 4 kyr before and lasted at least 1 kyr after the maximum event. Even though maximum extent most probably was attained at different times, the uncertainties in the data sets from the Kanin Peninsula, Kuloi and Pinega areas are too large to indicate any significant geographical difference in age. The new dates from the northern part of the Arkhangelsk district has not added significantly to previous estimates based on data restricted to the Severnaya Dvina area (Larsen *et al*. 1999). Also, exposure dating on the Kanin suggests that SIS began to retreat from the north-western part of Peninsula as early as 18.6 kyr and therefore neither contradicts this age estimate, nor provides a better accuracy. An average age of the maximum extent of the eastern part of Scandinavian Ice Sheet placed between 18 and 16 kyr is at least 3-5 kyr and maybe as much as 7-9 kyr younger than previously proposed by Arslanov *et al.* (1970, 1971).

Deglaciation drainage pattern

Palaeogeographic reconstructions depict the position of the retreating ice margin, the distribution of ice-dammed lake basins and the drainage ways of meltwater from the time of the largest extent of SIS (Fig. 8) until the Allerød interstadial (Figs 10A-D). Ice-dammed lake and river terraces levels are based on the present-day topography and geological mapping. The limitation of ages from melting ice facies only gives a rough estimate of the duration and spatial development of sedimentation during the deglaciation (Fig. 9). Ages of sediments and plant detritus from deglaciation basins indicate retreat of the active margin of SIS, downwasting, thermo-karst activity and immigration of a sub-arctic flora from *c.* 15 kyr and about five kyr onward.

Mezen Bay ice lake – Since the work of Ramsay (1911) many authors have suggested that the wide valley of the Kuloi River was the main discharge system of ice-dammed lakes either during last maximum of SIS or during the final stages of deglaciation (Yakovlev 1956). An ice-dammed lake developed in the Mezen Bay area (Figs 8, 10A; lake 1) and was fed from the melting ice sheet in the White Sea and from the Mezen and Pyoza Rivers draining ice free periglacial areas (Houmark-Nielsen *et al*. 2001). The level of the ice lake was about 15 metres a.s.l. and controlled by a threshold located on the lowermost part of the Kanin Peninsula on the waterdivide between the White Sea basin and the Chyosha Bay. Present-day valleys of the Chizha and Chyosha rivers (Fig. 1) crossing the peninsula are filled with 5-8 metres of fluvial sediments. The valley continues as a submarine canyon eastward at depths around 15-30 metre below sea level. It has a length of more than 20 km, the width is up to 1 km, and depth is up to 30 metres. The ice-dammed lake discharged to the Chyosha Bay. Laminated clay was formed on the bottom of this lake are located at altitudes of 5-17 metres a.s.l. (Kjær *et al.* 2003) and shown in the section from sites 13 and 15 (Figs 1, 3). The elevated river terraces observed by Zekkel (1939) that have altitudes of 5-6, 8-10, 20 metres a.s.l. on the lower Mezen and Pyoza rivers, may represent falling water level in the lake with time. A terrace at 15 metres at the Kanin Peninsula may correspond to one of these stages

Kuloi ice lake – This ice-dammed lake coexisted with an ice-dammed lake in the Mezen Bay until SIS lost contact with the Cape Kanin Nos which led to discharge of the lake onto the Barents Sea. The Kuloi ice lake filled the valley and the adjacent lowland of the Kuloi River

(Figs 8, 10A-B; lake 2). This lake was not more than 15-20 metres deep and it drained from a threshold in the northeast into the Mezen Bay ice lake. According to Zekkel (1939) the highest terraces of the Kuloi River is composed of coarse sand and has an altitude of 20-25 metres a.s.l. near the entrance to the lake. Combined with the mapping by Lavrov (1991) this indicates that the lake level was about 30 metres a.s.l. with a gradual drop to 16 metres. As deglaciation proceeded, the ice front retreated from the Lower Kuloi River which inevitably led to disappearance of the Kuloi ice-lake (Fig. 10B). When discharge from the Pinega and Dvina ice lakes occurred through the Kuloi River, a 3 km wide meltwater valley developed (Fig. 10 D).

Pinega ice lake – An ice-dammed lake occupied the middle Pinega River during the SIS maximum (Fig. 8; lake 3). The level of this lake was at 80 metres a.s.l. and controlled by a threshold on the water divide between the River Ezhuga and River Ezhuga Zyranskaya, with discharge eastward into the Mezen basin (Figs 8, 10A). These ancient ice-lake terraces with an altitude of 80-90 metres a.s.l. are known from both from the Severnaya Dvina and Pinega River lowlands (Atlasov *et al.* 1978; Arslanov *et al.* 1984). As the glacier retreated, the Severnaya Dvina ice-dammed lake merged with the Pinega ice lake (Fig. 10A), both discharging into the Mezen basin. Eventually the water level dropped from 80 to 50 metres a.s.l. and most Quaternary deposits were washed away from the banks of the Pinega River (Fig. 10B). The Pinega ice lake practically ceased to exist around 15$\pm$1.6 kyr, but the Pinega River maintained its discharge, now supplied with melt water from the Severnaya Dvina ice lake. As the ice melted further away, discharge carved a deep spillway at altitudes of 18-20 metres a.s.l. on the Kuloi River – Pinega River watershed (Fig. 10D). The age of events are based on dates of aeolian sediments lying at altitudes about 30 metres a.s.l. in the former lake basin at the Kulogora village on the Pinega-Kuloi waterdivide (Figs 1, 7; site18).

Severnaya Dvina ice-lake – During the maximum extent of SIS an ice-dammed lake filled the upper parts of the Severnaya Dvina River basin (Fig. 8; lake 4). The level of the lake reached elevations around 130-120 metres a.s.l. (Atlasov *et al*. 1978; Arslanov *et al*. 1984) and run off was directed southward into the Volga basin (Kvasov 1975; Lunkka *et al.* 2001). As deglaciation proceeded and the ice front melted into the middle *lower* Dvina basin, a spillway opened eastward and the lake level dropped to about 80-90 metres a.s.l. and discharge was redirected into the Mezen Bay through the middle part of Pinega and Mezen rivers (Fig. 10A).

When the ice melted away from the mouth of the Ezhuga River, the lake level dropped from 80 to 50 metres a.s.l. and a new 50 metre threshold developed on the water divide of the Pokshenga-Pukshenga rivers, while run off was directed into the Mezen River (Fig. 10B). Possibly this occurred about 15-15.7$\pm$1.2 kyr as indicated by dates of lacustrine deposits underlying subaerial sediments in the Chelmokhta sections at an altitude 12 metres a.s.l. (Fig. 1, site 24 and Figs 9, 11) (Larsen *et al*. 1999). At Raibola further up stream in the Dvina basin (Fig. 1; site 30), fluvial sand resting on laminated clays at an altitude of 56 metres a.s.l show OSL ages of 17.2 and 14.0 kyr (Larsen *et al*. 1999). We may conclude that the lake level had dropped from 80 to 50 metres a.s.l. in the Dvina basin about 15 kyr ago.

As the active ice margin retreated from the mouth of the Pinega River, the level dropped from 50 to 18 metres a.s.l. and water discharged through the lower Pinega River into the Kuloi River basin and the Dvina ice lake ceased to exist (Fig. 10C). This probably occurred about 13.7 kyr ago, as indicated by subaerial deposition at Chelmokhta (Fig. 1; site 24 and Figs 9, 11) (Larsen *et al*.1999). Later a new Severnaya Dvina ice lake appeared near the Mouth of present river at an altitude at 18-20 metres a.s.l (Fig. 10D). Discharge from the lake was into the Mezen Bay through the rivers Pinega and Kuloi, until the time, when the ice front retreated from the southern part of the Dvina Bay (Fig. 10D). This probably occurred immediately before the Allerød event, as indicated in Psaryovo section (Fig. 1; site 41), where plant remains with an age 13.2 kyr occur in fluvial sand with an altitude of 25 metres a.s.l. (Larsen *et al*.

1999). Pollen studies indicate Older Dryas and Allerød ages of varved and homogenous silt and clay known from boreholes north of Psaryovo section (Fig. 1; site 41) ranging from altitudes of 34 metres a.s.l. to zero (Baranovskaya *et al*. 1977; Pleshivtseva 1977).

Chelmokhta, an example of aerial downwasting – At Chelmokhta (Fig. 1; site 24) the ice locally moving from northerly directions ceased to flow and the development of constantly subsiding ice-confined depressions caused the sedimentation of a variety of supraglacial, mostly waterlain, sediments found up to about 20 metres above present river level (Fig. 11). The sediments are exposed in the river banks of the present day Dvina River 175 km upstream from Arkhangelsk. Seven sections along a 1 km long segment of the river were documented. The sedimentary successions were all influenced by mass-movement and gravitational sliding during inversion of the topography due to melting of stagnant ice from beneath (Lyså *et al.* 2001).

Overlying basal till with Scandinavian provenance sediments of four major depositional events can de detected. Lowermost, the basal till gradually changes into stratified diamict interbedded with sorted sand and mud showing upward decreasing clast content, all deposited in local depressions. In some sites imbricate gravel or cross bedded sand with an erosional lower contact suggest deposition in melt water streams, which possibly signal the lowering of the base level caused by drop in water table in the Dvina ice lake about 15 kyr ago. Laterally restricted and strongly deformed thin beds of sand and mud overlie these deposits. The sediments show normal as well as inverse grading, and combined with laminated mud they indicate deposition by sediment gravity flows and fall out from suspension. Sand could either have been blown by wind or transported by melt water into small ice dammed ponds and lakes. Plant remains such as lenses of peat, twigs and leaves indicate the development of a vegetation cover on the clastic debris covering the stagnant ice. Shells of *Anodonta* indicate the immigration fresh water molluscs into the lakes. The flora counted species like *Betula, Salix, Populus, Larix* and *Picea* (Lyså *et al.* 2001).

Finally, deposition of rhythmically laminated mud with occasional diamict lenses, drop stones and beds of sand was followed by an episode of erosion caused by sudden lowering of the local water level. This is indicated by deposition of thick sheets of bedded flow diamict and massive sand containing boulders, rip-up clasts of peat and bedded lacustrine sand with leaves and twigs. Because plant remains from the underlying units range in age between 11.8 and 10.5 cal ^{14}C kyr (Larsen *et al.* 1999), we estimate that the final collapse of the dead ice took place in the Early Holocene. Possibly at the same time the White Sea was inundated by arctic marine waters. Radiocarbon ages of 9330$\pm$120 yr BP on marine shells indicate that the ocean waters had inundated the White Sea around 11 kyr ago (Koshechkin *et al*. 1977) and subsequent regression began at the beginning of the Boreal period about 9100 yr BP (Baranovskaya *et al*. 1977).

Ice sheet dynamics and flow pattern

The ice sheet configuration and flow pattern along the eastern flank of the last SIS in northwest Russia is illustrated in Figure 12. The SIS invaded southern and central Finland less the 25 kyr ago, as indicated by dating of mammoth tusks which pre-dates deposition of till by the last SIS (Lunkka *et al.* 2001; Ukkonen *et al.* 1999). In the following 5-9 kyr, the glacier front had advanced with an average of at least 10-12 km per hundred years about 500-800 km towards the east and southeast. The maximal position in northwest Russia was reached at 19-15 kyr ago (Fig. 12). Ice sheet flow through large water filled bedrock depressions on the rim of the Baltic shield and the Palaeozoic platform to the east probably enhanced velocities. Corridors of rapid flowing ice separated by slower flowing ice constituted ice divide zones.

Our results suggest that glacier tongues and lobes, probably less than 300 metres in thickness, not only at the maximum glaciation itself, but also during different phases of deglaciation protruded the general north south trending ice margin and flowed rapidly eastward through shallow depressions on the low relief northwest Russian plain. This hypothesis, which remains to be thoroughly tested, is based on theoretic models of similar dynamic behaviour to that of terrestrial based ice streams as predicted by Stokes and Clark (1999) and Boulton *et al.* (2001). Previously, Lagerlund (1987) had introduced the concept of ´outlet surges´ in which minor depressions in the overall pre ice-advance topography of the circum Baltic lowlands determined the distribution of proglacial lakes, whose water bodies and sediments acted at gateways for rapid flowing ice. Similar pre-requisites including little or no topographic control and special bed lithologies for the location of potential fast ice flow corridors have recently been revived within the Laurentian Ice Sheet (Stokes & Clark 2003).

In areas along the periphery of SIS elsewhere, former terrestrially based ice streams had supposedly been operating (Clark *et al.* 2003). These areas coincide with a region of sedimentary bedrock and thick Quaternary deposits surrounding the crystalline Fennoscandian Shield. The largest extent of the SIS was reached earlier and under colder climatic conditions in southwestern Scandinavia and the Baltic compared to northwestern Russia. Because similar glaciation scenarios including fast and canalised ice flow occurred around the whole SIS, factors like the presence of water filled basins and deformable substratum must have played a major role. From the northern North Sea and to the circum Baltic region rapid flowing ice streams bounded by slow flowing inter stream areas dominate the pattern of Late Weichselian glaciation. Except for the Main Glaciation interstream-event in Denmark (22-19 kyr), the Norwegian Channel Ice Stream (26 kyr; Sejrup *et al.* 1994) and, the Kattegat Ice Stream (29-25 kyr) and the Young Baltic Ice Streams (19-15 kyr) all advanced from SIS across marine waters or ice dammed lakes. These advances show flow patterns and left behind morphological features similar to that of terrestrial based ice streams (Boulton *et al.* 2001; Clark *et al.* 2003 and references herein; Houmark-Nielsen & Kjær 2003).

We therefore propose that also in northwest Russia ice dammed lakes and water saturated fine grained proglacial sediments in topographic depressions could have caused rapid outlets from the periphery of an assumingly quite steep sloped margin of the main ice sheet. Moreover, glacier flow velocities were probably reduced and the ice considerable thinner in areas of elevated Pre-Quaternary bedrock. In these slow flowing inter-lobe areas third order flow divides *sensu* Boulton *et al.* (2001) could have been located. Along the north-eastern part of the ice sheet in Russia, the Barents Sea - Kanin fast flowing ice moved from the west and north-west and rimmed the ice divide zone located on the central part of the Kola Peninsula (Fig. 12; I). The ice occupied the northern part of the White Sea and reached the northwestern point of the Kanin Peninsula where the Scandinavian Ice Sheet merged with the Barents Sea Ice Sheet.

South of Kola Peninsula, the White Sea ice flowed from the west along the White Sea depression and dispersed into a number of ice lobes in the Arkhangelsk region (Fig. 12; II). The Kuloi ice tongue (Fig. 12; IIA) moved to the northeast along the White Sea inlet. It was flanked to the north by the Kola Peninsula highland and to the east and south by the Kuloi Plateau where it reached heights of 150-220 metres. The Dvina and Pinega ice tongues (Fig. 12; IIB, IIC) flowed from the NW along Severnaya Dvina and Pinega lowlands along with fast flowing ice in the Onega River depression (Fig. 12; IID), which also had its origin in the White Sea. Further southward other corridors of rapid ice flow in the SIS moved from more northerly directions and occupied the depressions and adjacent areas of the lakes Onega (III), Ladoga (IV) and Chudskoe (V). Ice divides were situated along bedrock highs of 150-300 metres (Aseev 1974).

Decay of the ice sheet

Event-stratigraphic models in the region often relate Late glacial climatic variations with the ice-marginal positions and recessive stages of the SIS (Aseev 1974; Krasnov 1971), but input to our model does not allow challenging this hypothesis. The sediment successions at Chelmokhta (Fig. 11) indicate deposition in an environment of aerial downwasting of buried stagnant glacier ice that was subjected to thermo-karst processes, the ice along the lower-middle Dvina basin had been detached from the active ice sheet. Subsidence and sagging caused the development of kettle holes with accommodation space for deposition of a variety of sedimentary facies and the re-sedimentation of unstable soils and remnants of the pioneer vegetation by gravitational slumping and sliding. The sub-arctic flora arrived in the region at the beginning of the Allerød period (Baranovskaya *et al*. 1977; Wohlfarth *et al*. 2002, 2004) while buried glacier ice was melting causing inversion of the landscape as was also the case in the area northeast of the Arkhangelsk district which was covered by an older Weichselian ice sheet from the Barents-Kara Seas (Tveranger *et al*. 1995). From the eastern North Sea across the southern Baltic to northwest Russia Late Weichselian streamlined terrains separated by belts of terminal moraines possibly generated by narrow and rapid flowing ice lobes are overprinted by landforms which relate to aerial downwasting (c.f. Boulton *et al*. 2001). In Russia an inter Bølling-Allerød terminal belt (Neva stage) separates a region of dead ice relief generated by downwasting to the east and to the south and another with evidence of frontal deglaciation to the west and north (Fig. 13). The terminal zone stretches from the Baltic Sea south of the Gulf of Finland via the southern shores of Lake Ladoga and through the lake Onega into the Onega Bay and further out into the White Sea. Further north the terminal belt encircled the highland of the Kola Peninsula. Between the maximum position of SIS and the Neva stage termini a more than a several hundred thousand km^2 large zone of downwasting contains a complex dead ice relief, including fields of kame-like hills and hummocky moraine. Over large area glaciolacustrine sediments may blur the original shape and size of the dead ice landforms.

We suppose that long and thin lobes of fast flowing ice were detached from the main ice sheet once the ice sheet profile along the flow paths had become flat enough. Thus, the ice lobes were no longer fed by the ice sheet and the flow ceased. The marginal zone now became subjected to the melting of large dead ice masses. In northwest Russia detachment of peripheral debris rich lobes from the flowing ice began *c.* 16-15 kyr and lasted maybe until the beginning of the Allerød Interstadial (Fig. 13). Bølling sediments hardly occur in this zone probably because the larger part of area was covered by unstable dead-ice. The age of the oldest organic remnants in sediments from lakes across the downwasting zone belong to the Allerød Interstadial (Ekman & Iljin 1995, Davydova *et al.* 1998). Deposition of organic debris in small lakes on the watersheds covered by dead ice began as late as Preboreal and Boreal (Demidov & Lavrova 2001; Wohlfarth *et al.* 2002, 2004). Thus huge areas in the peripheral parts of SIS stagnated rapidly, fields of debris covered dead ice spent 4-7 kyr of degradation as the climate was cold and permafrost still prevailed and because by thick accumulations insulating overburden retarded downward directed heat transfer delaying the downwasting.

From the Allerød Interstadial and onwards, decay of the active ice margin changed to the frontal type deglaciation. The ice sheet rested on crystalline bedrock and was not to a similar degree enriched with debris, as had been the case further east where it flowed over deformable sediments. The relatively thin and lobe shaped ice front quickly melted during the ameliorated climate of the Allerød oscillation. The glacier retreated about 150-200 km from the Neva terminal belt to the Young Dryas margin in *c.* 800 years (Fig. 13). Supraglacial sediments rarely occur in this area, but fields of drumlins are widespread as are long and prominent eskers systems (Ekman & Iljin 1995). Under both types of deglaciation, the large ice-dammed lakes in the Baltic, the White Sea, and the Ladoga and Onega Lakes assisted calving and increased the rate of deglaciation.

Conclusions

Till deposited by the last SIS (Bobrovo till) is identified by its stratigraphic position, its contents of Scandinavian erratics and its directional properties indicating ice flow from westerly directions.

The eastern flank of the Late Weichselian Scandinavian Ice Sheet reached its maximal position in northwest Russia sometime between 20 and 15 ka BP, in the Arkhangelsk region this probably occurred around 18-16 ka ago. This position was reached up to 10 kyr after maximum extent in the south-western part of SIS.

At the maximum, the ice margin stretched from the NW part of the Kanin Peninsula across bottom of the White Sea southward to the mouth of Kuloi River. It touched the eastern part of the Kuloi Plateau and covered the lower and middle parts of the Pinega River and turned westward into the Severnaya Dvina River lowland.

Fast flowing ice lobes along the fringes of the last Scandinavian Ice Sheet were guided by deformable sediments and ice-dammed lakes. Lobes were detached and isolated from the main ice body when reaching the maximum extent. The glaciers stagnated into large dead ice fields and began to melt about 15-16 kyr ago.

In the zone between the former margin of SIS and the Neva terminal belt aerial down wasting caused the inversion of landscapes. Deglaciation sediments are composed of sediment gravity flows of diamicts material, lacustrine mud with plant debris and aeolian and fluvial sand, all often strongly deformed by slumping, sliding and periglacial frost-thaw processes. Deposition lasted until the beginning of the Holocene and was accompanied by the slow melting of buried and stagnant ice and the migration of a sub-arctic flora and fauna.

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Figure captions

Fig.1

The topography of the Arkhangelsk Region, northwest Russia. Map shows alternative configurations of the maximum extent of the last Scandinavian Ice Sheet. Numbers 1- 48 refer to investigated sections and other localities in the region. 1- Tarkhanov, 2 - Krynka, 3 - Madakha, 4 - Ostraya, 5- Pestsovaya, 6 - Konushin, 7 - Konushinskaya Korga, 8 – Morzhovets Island, 9 – Cape Kargovsky, 10 – Cape Abramovsky, 11 - Koida, 12 - Megra, 13 – Cape Tolstic, 14 & 15 – Syomzha 1&2, 16 - Zaton, 17 - Bychie, 18 - Kulogora, 19 - Ezhuga, 20 - Karpogory, 21 - Shilega, 22 -Verkola, 23 - Olema, 24 - Chelmokhta, 25 - Lipovik, 26 - Tomasha, 27 - Erga, 28 - Boltinskaya, 29 - Yumizh, 30 -Raibola, 31 - Smotrakovka, 32 - Osinovskaya, 33 - Ust-Padenga, 34 - Koleshka, 35 - Pasva, 36 - Nyandoma, 37 - Mosha, 38 - Telza, 39- Bobrovo, 40 - Trepuzovo, 41 - Psaryovo, 42 - Chavanga, 43 - Strelnya, 44 - Kumzhevaya, 45 - Babia, 46 - Kachkovka, 47 - Iokanga, 48 - Oiva, 49 - Tova. Compiled from the present study, Devyatova 1969, 1982; Armand 1969 (Kola peninsula); Larsen *et al.* 1999, Atlasov *et al.* 1978, Arslanov *et al.* 1984 (Dvina lowland) and Lunkka *et al.* 2001.

Fig. 2

The main Quaternary geomorphological features of the Arkhangelsk Region and the position of the largest extent of the Scandinavian Ice Sheet in the Late Weichselian. Compiled from Krasnov (1971), Ganeshin *et al.* (1980) and Lavrov (1991)

Fig. 3

Stratigraphic logs from selected sections along the northern shores of the Kuloi Plateau.

Fig. 4

Clast fabric diagrams from selected sections.

Fig. 5

Sketch of the section at Tarkhanov, northwest shores of the Kanin Peninsula. Late Pleistocene deposits are banked upon the bedrock cliff. Luminescence dates of sediments and cosmogenic exposure dates on bedrock are indicated. Inferred glacier limit is shown.

Fig. 6

Till with Scandinavian clasts and boulders underlying the deflation surface, Tarkhanov section, Kanin Peninsula.

Fig. 7

Stratigraphic logs from selected sections on the middle parts of the Pinega River.

Fig. 8

Palaeogeographic reconstruction of the eastern flank of Scandinavian Ice Sheet at its maximum extent. Numbers refer to ice dammed lakes, 1: Mezen Bay ice lake, 2: Kuloi ice lake, 3: Pinega ice lake, 4: Severnaya Dvina ice lake.

Fig. 9

Plot of luminescens, radiocarbon and cosmogenic exposure ages, northwest Russia (one standard derivation). Dated sediments and plant remnants are from below and above Bobrovo till deposited by the last Scandinavian Ice Sheet. Dated extramarginal deposits relate to phases of ice damming or deglaciation. Compiled from the present study, Larsen *et al.* (1999); Lunkka *et al.* (2001) and Kjær *et al*. (2003)

Fig. 10

Palaeogeographic reconstructions picturing four stages (A-D) of Lateglacial ice retreat, deglaciation, development of ice-dammed lakes and run off patterns in northwest Russia.

Fig. 11

Stratigraphic logs from seven sections at Chelmokhta, Severnaya Dvina River (Fig. 1; site 24). Sediments above Bobrovo till indicate aerial downwasting, inversion of landscape and deposition under thermokarst conditions. Luminescence and calibrated ^{14}C ages are from Larsen *et al.* 1999.

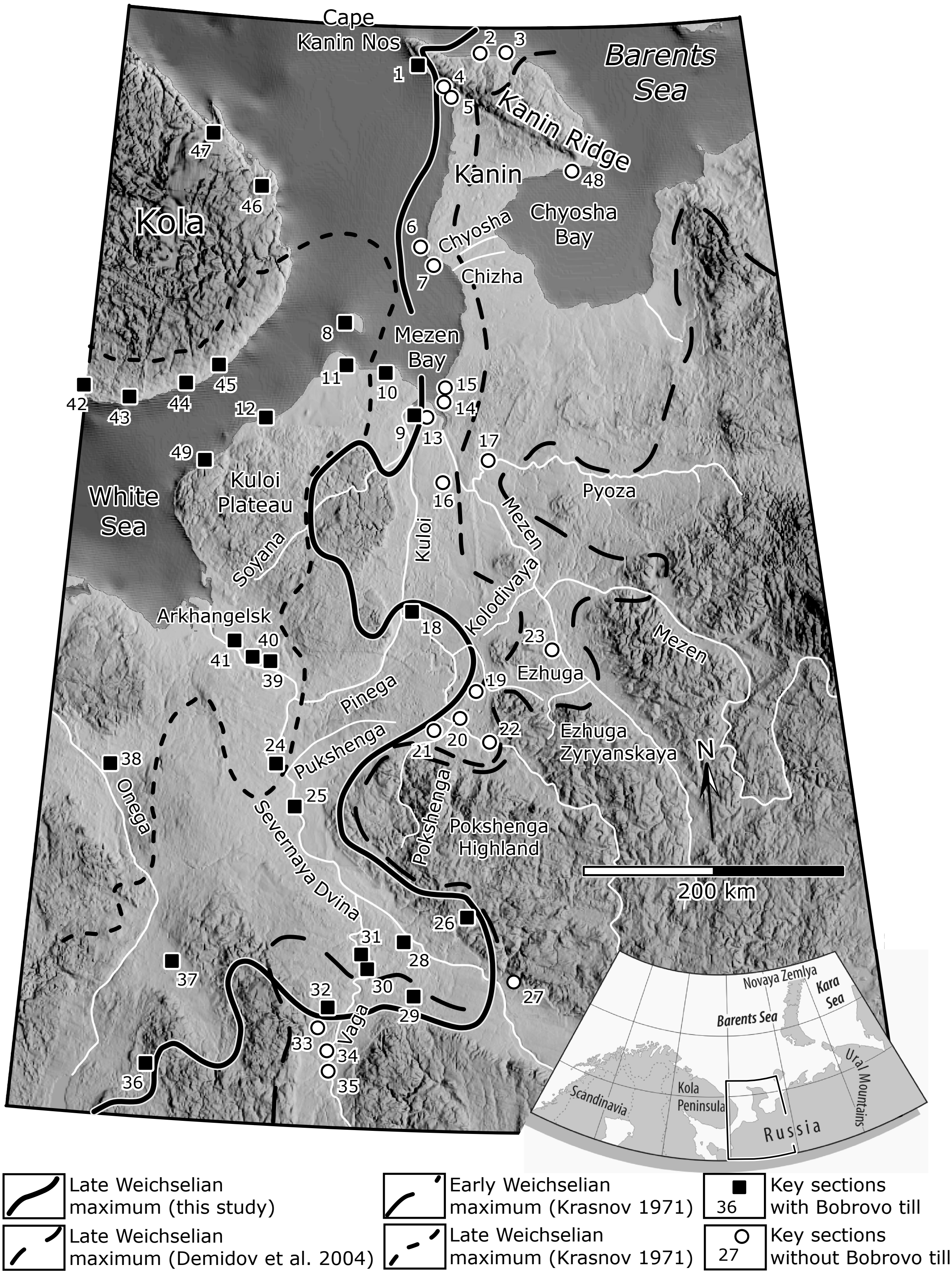
Fig. 12

Reconstruction of glacier flow patterns, marginal position and ice dammed lakes during the maximum extent of the eastern flank of the Scandinavian Ice Sheet 18-17 ka BP. Numbers I-V indicate the corridors of fast flowing ice. Compiled from the present study, Aseev (1974), Gey & Malakhvsky (1998), Ekman & Iljin (1995) and Lunkka *et al* (2001).

Fig. 13

Reconstruction of the degradation along the eastern flank of the last Scandinavian Ice Sheet. Vast regions of aerial downwasting are indicated east of the Bølling-Allerød ice marginal position.

Data from the present study is compiled with data from Aseev 1974, Gey & Malakhvsky 1998, Ekman & Iljin 1995, Lunkka *et al.* 2001, Hättestrand 2006.



Cape
Kanin Nos
Barents
Sea
Kanin Ridge
Kanin
Chyosha
Bay
Chyosha
Chizha
Kola
Mezen
Bay
White
Sea
Kuloi
Plateau
Pyoza
Mezen
Kuloi
Soyana
Kolodivaya
Arkhangelsk
Ezhuga
Mezen
Pinega
Pukshenga
Ezhuga
Zyryanskaya
N
Onega
Severnaya Dvina
Pokshenga
Pokshenga
Highland
200 km
Vaga
1
2
3
4
5
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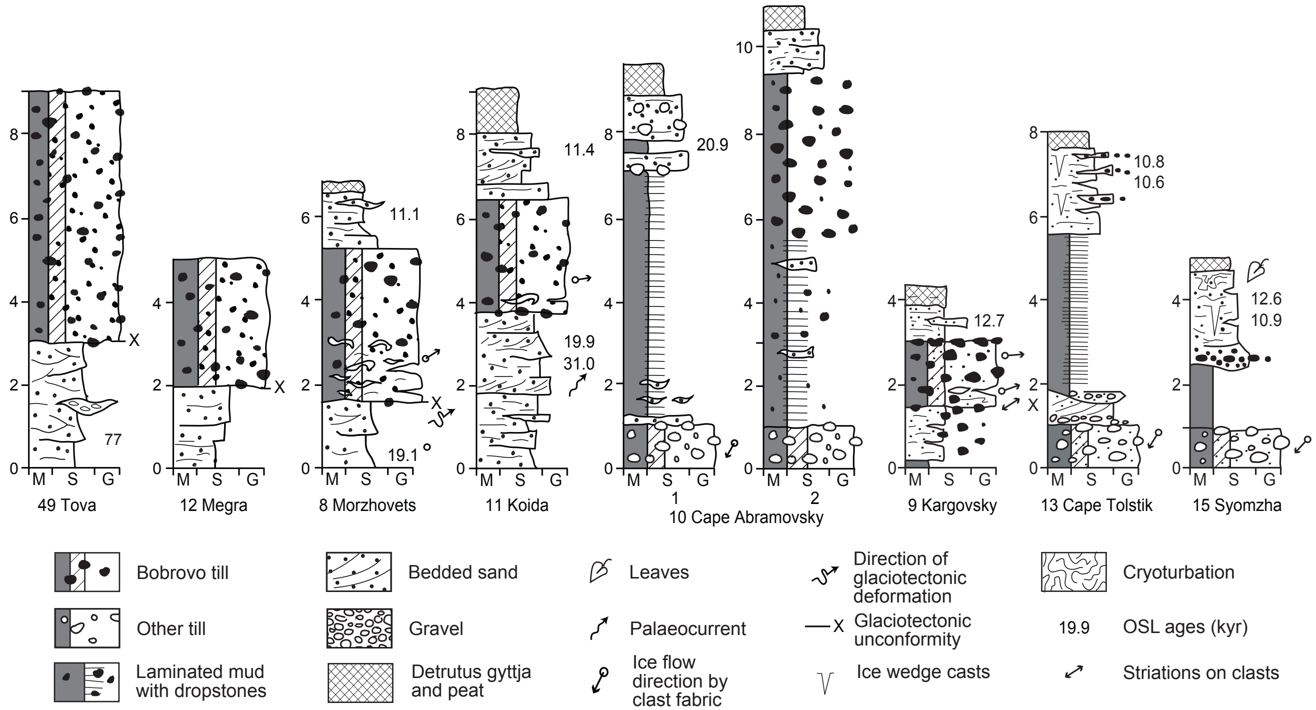
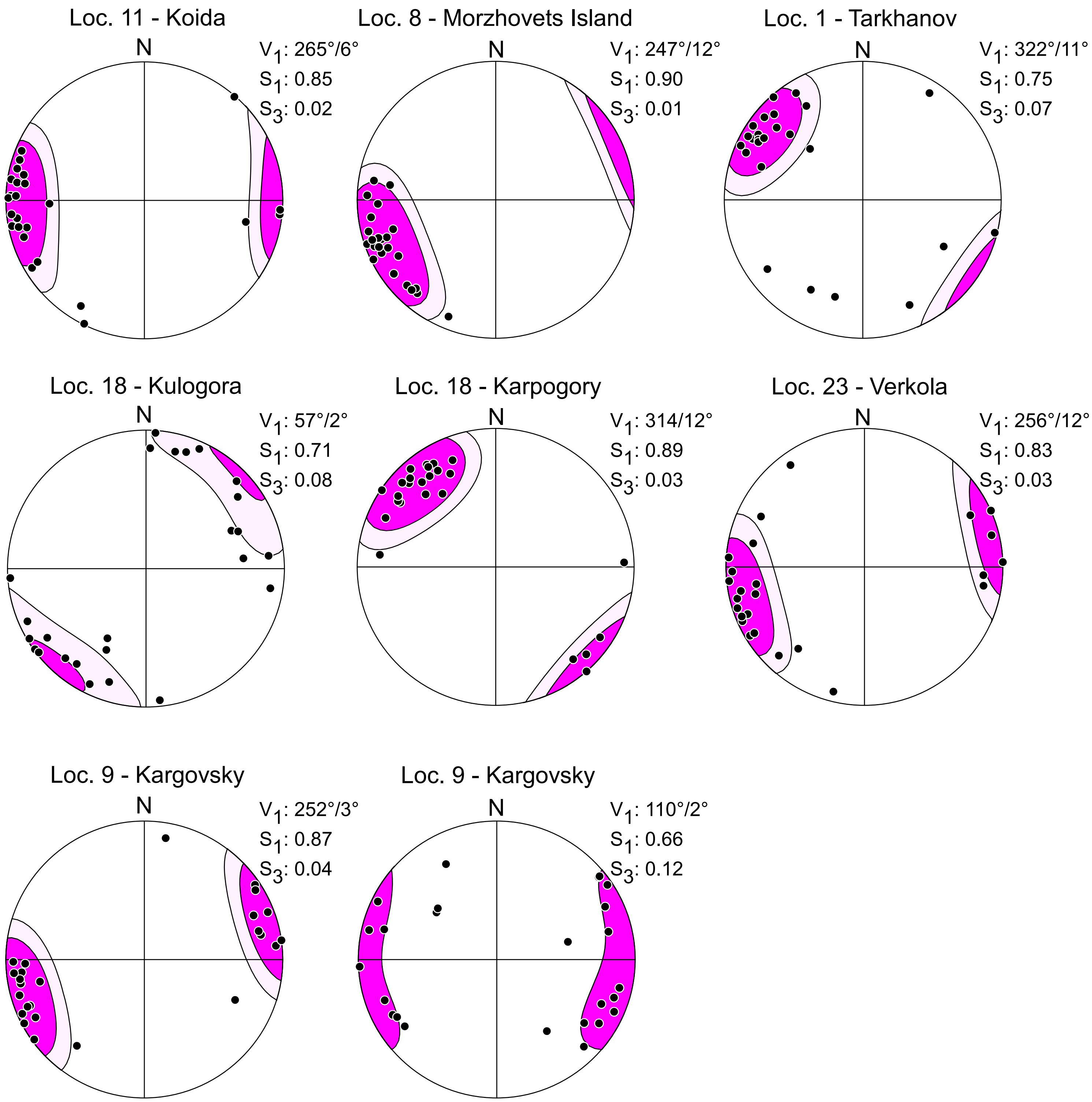
Novaya Zemlya
Kara
Sea
Barents Sea
Ural Mountains
Scandinavia
Kola
Peninsula
Russia
Late Weichselian
maximum (this study)
Early Weichselian
maximum (Krasnov 1971)
36
Key sections
with Bobrovo till
Late Weichselian
maximum (Demidov et al. 2004)
Late Weichselian
maximum (Krasnov 1971)
27
Key sections
without Bobrovo till

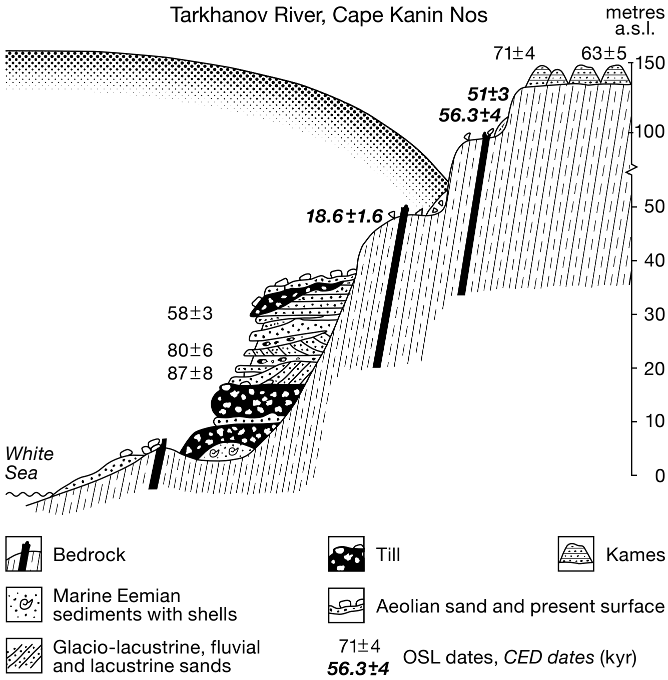

40km
Barents Sea
Kola Peninsula
Kanin Peninsula
White Sea
Maximum extent of SIS
End moraines and hummocy moraines
Kames
Sandar
Glaciolacustrine sediments
Weathered bedrock
Escarpments and slopes
Valley
Plateau
Channel

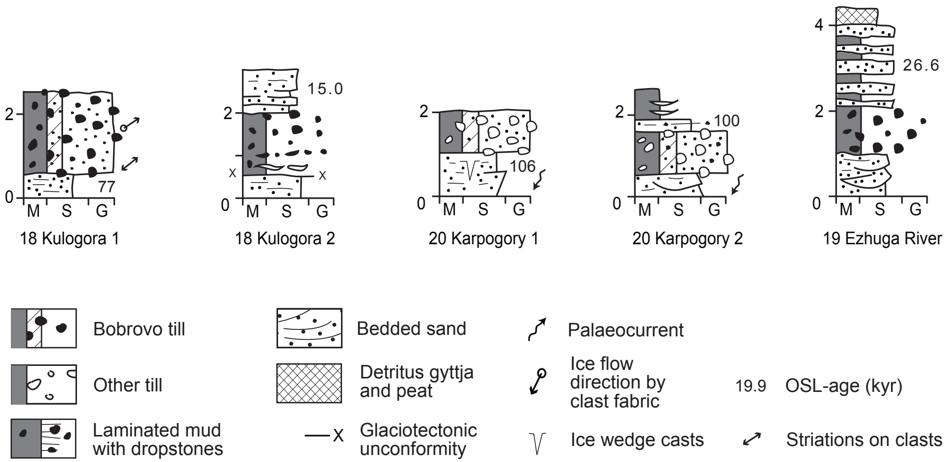


houmark michael /igor / fig3

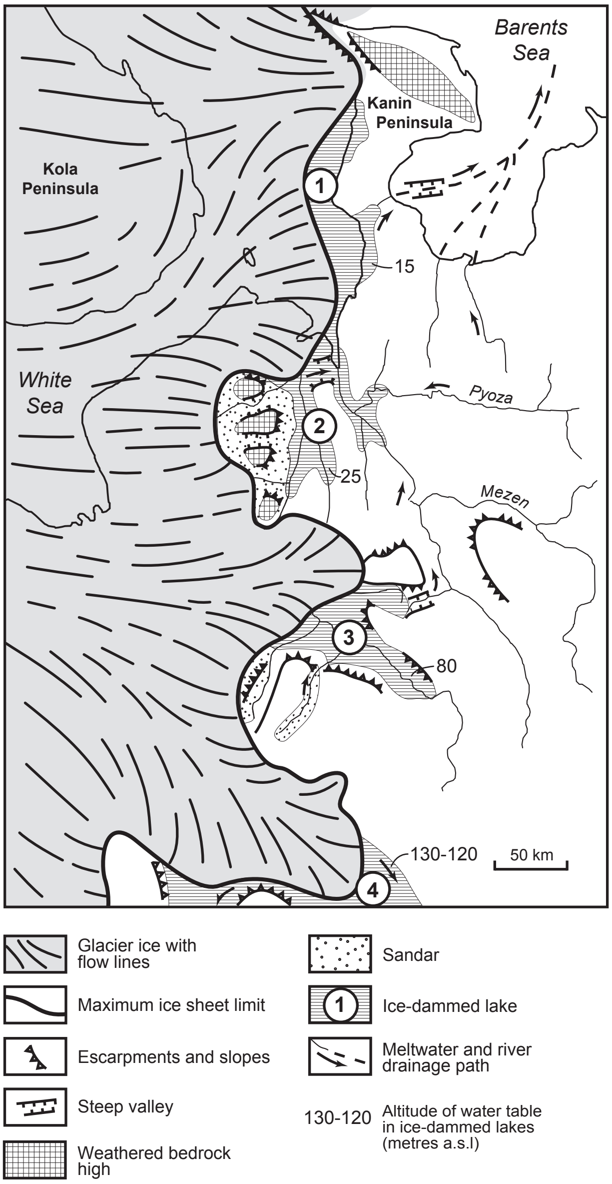


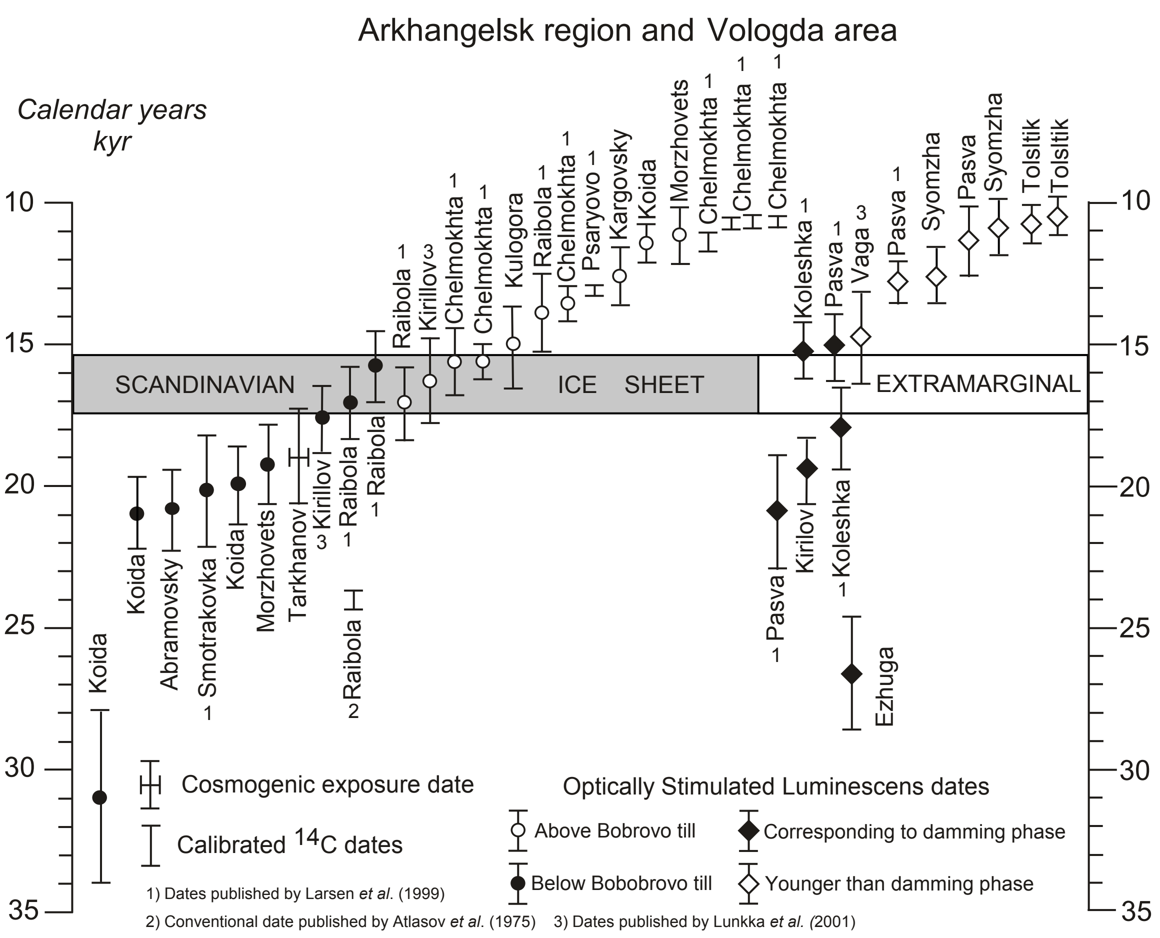
Loc. 11 - Koida
N
V1: 265°/6°
S1: 0.85
S3: 0.02
Loc. 8 - Morzhovets Island
N
V1: 247°/12°
S1: 0.90
S3: 0.01
Loc. 1 - Tarkhanov
N
V1: 322°/11°
S1: 0.75
S3: 0.07
Loc. 18 - Kulogora
N
V1: 57°/2°
S1: 0.71
S3: 0.08
Loc. 18 - Karpogory
N
V1: 314/12°
S1: 0.89
S3: 0.03
Loc. 23 - Verkola
N
V1: 256°/12°
S1: 0.83
S3: 0.03
Loc. 9 - Kargovsky
N
V1: 252°/3°
S1: 0.87
S3: 0.04
Loc. 9 - Kargovsky
N
V1: 110°/2°
S1: 0.66
S3: 0.12


Tarkhanov River, Cape Kanin Nos
metres
a.s.l.
71±4
63±5
51±3
56.3±4
18.6±1.6
58±3
80±6
87±8
White
Sea
150
100
50
40
30
20
10
0
Bedrock
Till
Kames
Marine Eemian
sediments with shells
Aeolian sand and present surface
Glacio-lacustrine, fluvial
and lacustrine sands
71±4
56.3±4
OSL dates, CED dates (kyr)

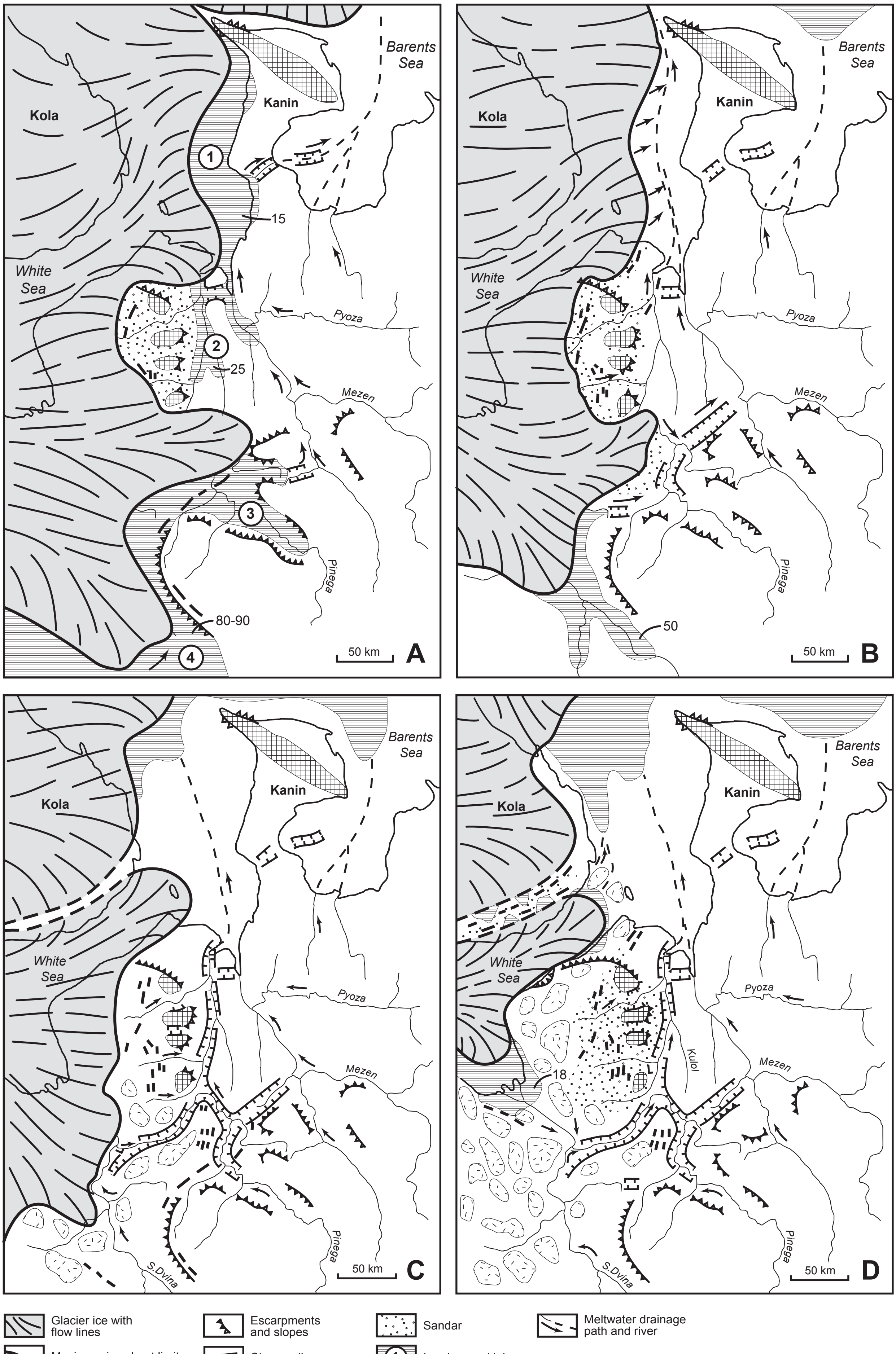


houmark michael /igor / fig7

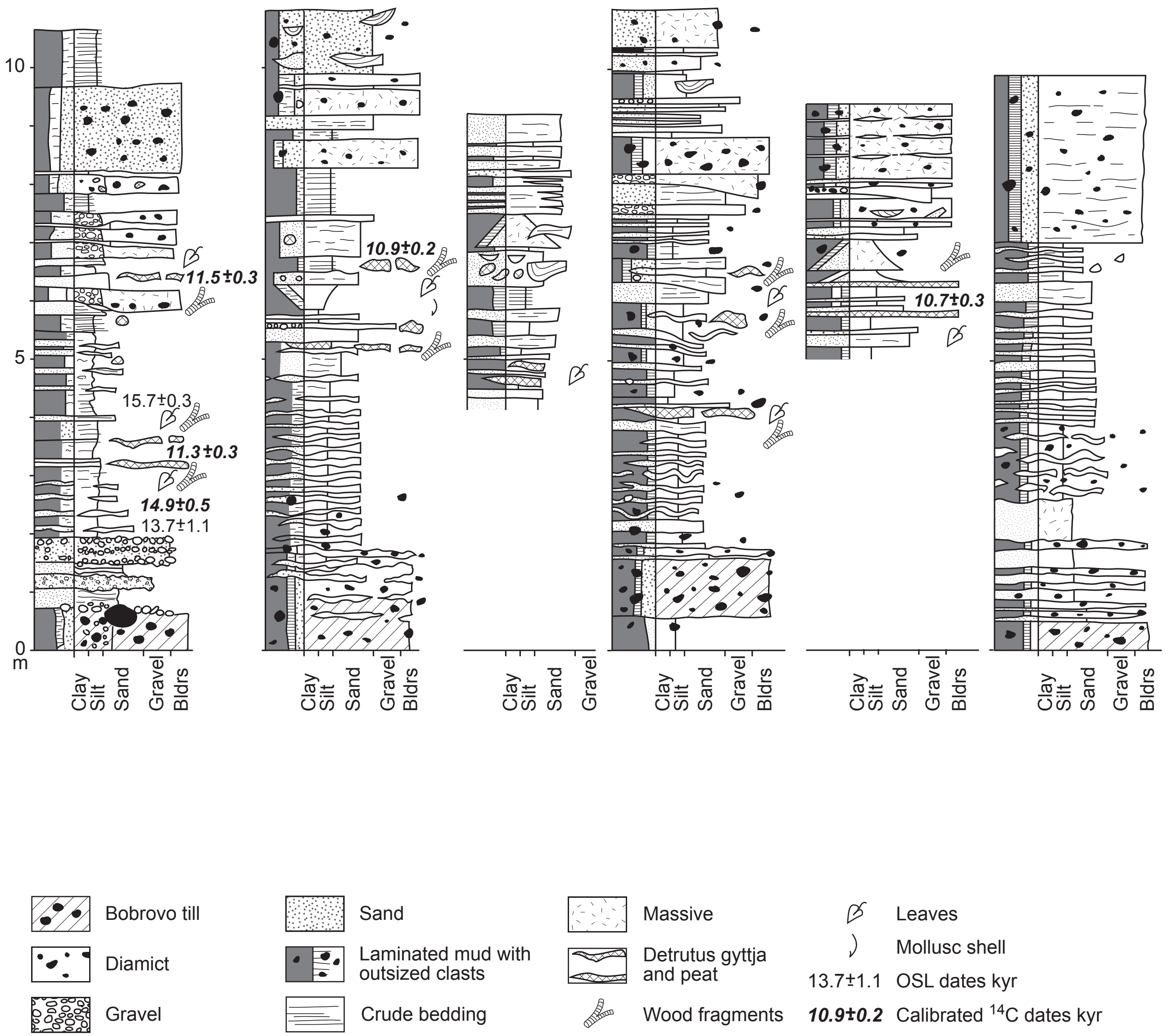

Barents Sea
Kanin Peninsula
Kola Peninsula
White Sea
Pyoza
Mezen
1
2
3
4
15
25
80
130-120
50 km
Glacier ice with flow lines
Maximum ice sheet limit
Escarpments and slopes
Steep valley
Weathered bedrock high
Sandar
Ice-dammed lake
Meltwater and river drainage path
130-120 Altitude of water table in ice-dammed lakes (metres a.s.l)



Arkhangelsk region and Vologda area
Calendar years kyr
10
15
20
25
30
35
SCANDINAVIAN
ICE SHEET
EXTRAMARGINAL
Koida
Koida
Abramovsky
Smotrakovka 1
Koida
Morzhovets
Tarkhanov
Kirillov 3
Raibola 1
Raibola 1
Raibola 2
Raibola 1
Kirillov 3
Chelmokhta 1
Chelmokhta 1
Kulogora
Raibola 1
Chelmokhta 1
Psaryovo 1
Kargovsky
Koida
Morzhovets
Chelmokhta 1
Chelmokhta 1
Chelmokhta 1
Koleshka 1
Pasva 1
Vaga 3
Pasva 1
Syomzha
Pasva
Syomzha
Tolstik
Tolstik
Pasva 1
Kirilov
Koleshka 1
Ezhuga
Cosmogenic exposure date
Calibrated 14C dates
Optically Stimulated Luminescens dates
Above Bobrovo till
Below Bobobrovo till
Corresponding to damming phase
Younger than damming phase
1) Dates published by Larsen et al. (1999)
2) Conventional date published by Atlasov et al. (1975)
3) Dates published by Lunkka et al. (2001)



houmark michael / igor / fig10



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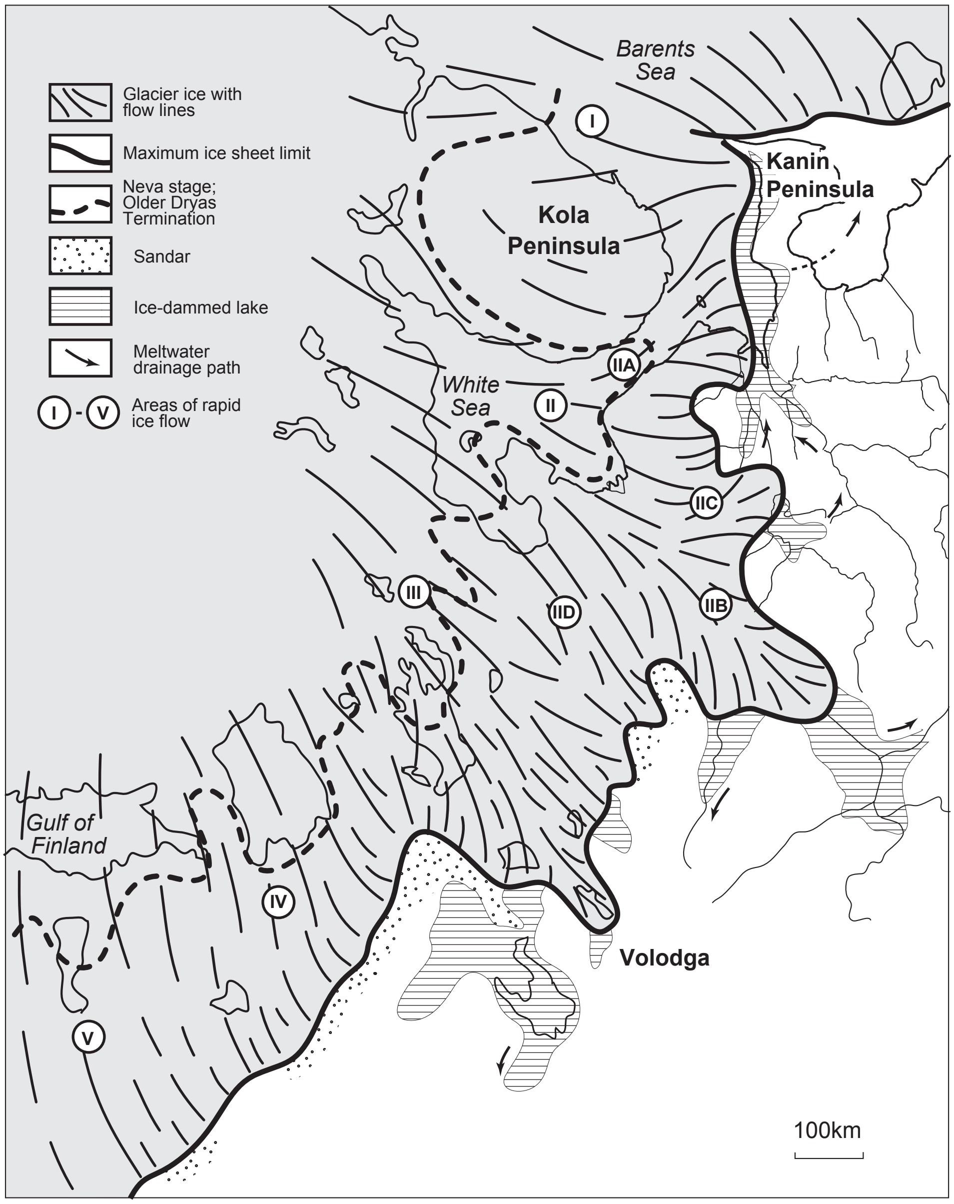

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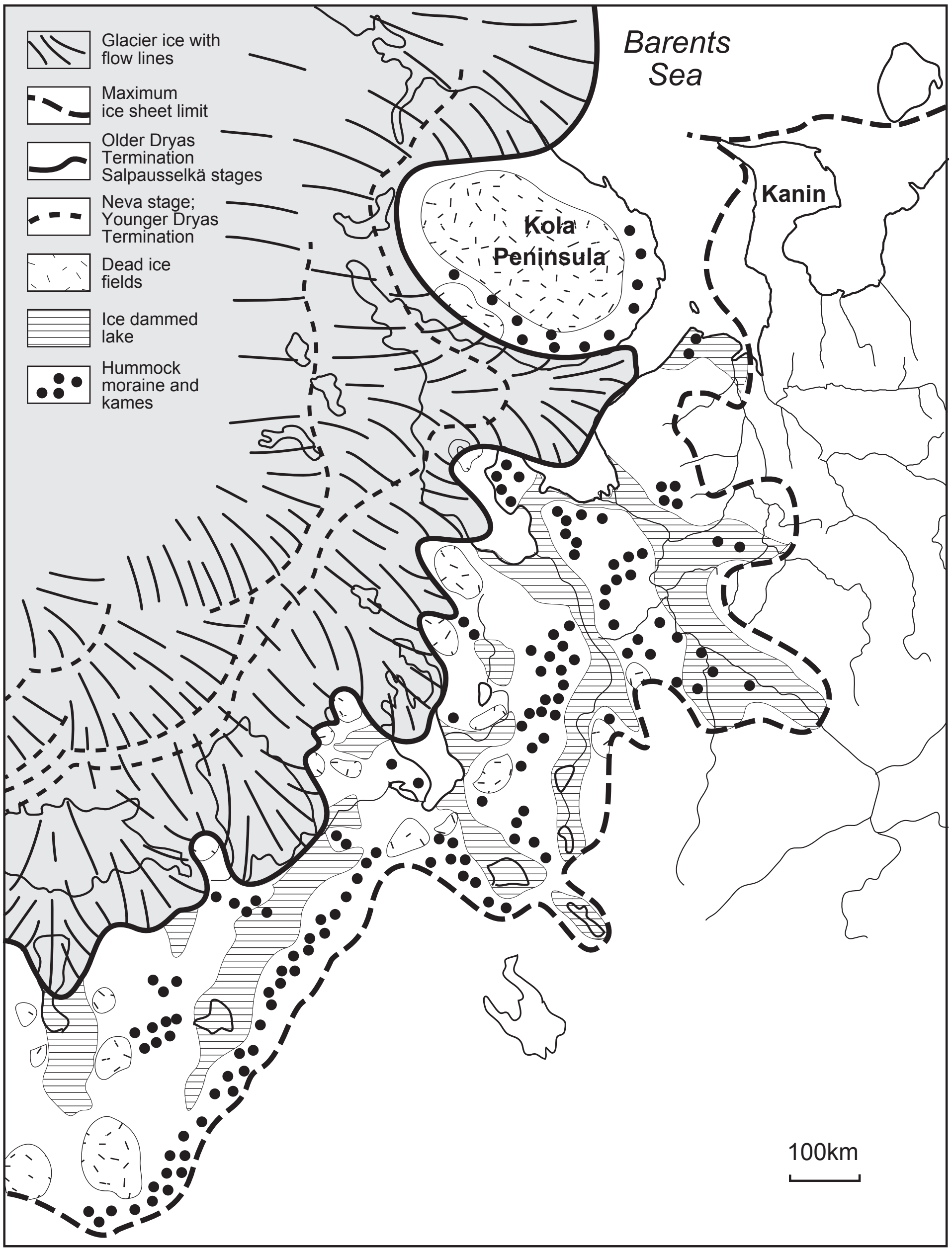

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Table 1. Luminescence (OSL) and cosmogenic exposure dates (CED) from the eastern part of White Sea and central Pinega River, northwestern Russia.

No. Fig. 1/ Strat. Pos.	Locality	Risø code	Sample No.	Age (ka)	Dated material	Dose rate	W.C. %
1/Below	Tarkhanov river	001020	00421	80±6	Glaciolacustrine	1.78	23
1/Below	Tarkhanov river	001021	00423	87±8	Glaciolacustrine	1.89	25
1/Beyond	Tarkhanov river	011001	01403	71±4	Glaciolacustrine	1.37	27
1/Below	Tarkhanov river	011002	01405	58±3	Glaciolacustrine	1.62	31
1/Beyond	Tarkhanov river	011030	01404	63±5	Glaciolacustrine	1.52	32
4/Beyond	Ostraya Hill	011014	01508	50±4	Lacustrine sand	2.44	30
5/Beyond	Pestsovaya Hill	011006	01500	72±4	Fluvial sand	1.66	33
6/Beyond	Cape Konushin	001015	00-410	94±6	Glaciofluvial	1.29	24
7/Beyond	Konushinskaya Korga	001011	00-406	66±7	Glaciolacustrine sand	2.26	11
8/Below	Morzhovets island	001035	00441	19.1±1.5	Glaciofluvial sand	0.83	20
8/Above	Morzhovets island	001036	00442	11.1±0.9	Glaciofluvial sand	1.38	21
9/Above	Cape Kargovsky	993828	99422	12.7±1	Fluvial sand	1.81	14
10/Below	Cape Abramovsky	001031	00437	20.9±1.7	Fluvial sand	1.19	27
11/Above	Koida river	001033	00439	11.4±0.7	Lacustrine sand	1.81	17
11/Below	Koida river	001032	00438	31±3	Fluvial sand	1.15	27
11/Below	Koida river	001034	00440	19.9±1.4	Fluvial sand	1.27	23
49/Below	Tova river	001037	00444	77±8	Marine sand	0.91	28
13/Above	Cape Tolstik	993817	99410	10.6±0.7	Fluvial sand	1.50	18
13/Above	Cape Tolstik	993818	99411	10.8±0.8	Fluvial sand	1.43	15
15/Above	Syomzha	001038	99404	10.9±0.8	Fluvial sand	1.90	18
15/Above	Syomzha	001039	99405	12.6±1.0	Fluvial sand	1.72	19
18/Below	Kulogora	993832	99426	77±7	Glaciofluvial sand	1.40	18
18/Above	Kulogora	993833	99427	15.0±1.6	Aeolian sand	1.21	18
19/Beyond	Ezhuga river	001043	99429	26.6±1.9	Glaciolacustrine sand	1.11	18
20/Beyond	Karpogory	993838	99432	106±8	Glaciofluvial sand	0.98	22
20/Beyond	Karpogory	993839	99433	100±7	Glaciolacustrine sand	1.72	12
Cosmogenic exposure dates							
1/Beyond	Tarkhanov river		00418	51.5±3	Quarts vein		
1/Beyond	Tarkhanov river		00419	55.0±4	Quarts vein		
1/Above	Tarkhanov river		00420	18.0±1.6	Quarts vein		