The last Scandinavian Ice Sheet in northwestern Russia: ice flow patterns and decay dynamics

IGOR N. DEMIDOV, MICHAEL HOUMARK-NIELSEN, KURT H. KJÆR AND EILIV LARSEN


Advance of the Late Weichselian (Valdai) Scandinavian Ice Sheet (SIS) in northwestern 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 BP 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. Ice-sheet collapse and deglaciation in the region commenced when ice lobes were detached from the main ice sheet. During the Lateglacial warming, disintegration and melting took place in a 200–600 km wide zone along the northeastern 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 subarctic flora invaded the region.

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The eastern flank of the Scandinavian Ice Sheet (SIS) occupied large parts of the Arkhangelsk district of the Russian European North during the Late Weichselian (Valdai). 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 are still subjects of debate (Svendsen et al. 2004). The duration and age of the glacial maximum are significantly different compared with the ice sheet’s termination nearer to the Atlantic continental margin. It therefore appears that the peak glaciation in Russia was delayed by almost 10 kyr. In southwest Scandinavia this state was reached some time between 32 and 25 kyr BP (Sejrup et al. 1994, 2003; Houmark-Nielsen & Kjær 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 BP (Svendsen et al. 2004, and references therein). The exact location of the merging is less evident, largely due to a 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 have until recently been poorly documented (Demidov et al. 2004).

In this article, 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 northwestern 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 river banks in the Arkhangelsk Region. By addressing these issues we close a gap in the
knowledge about the Late Weichselian glaciation history along the fringes of the entire 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) rugged denudation plain composed of Palaeozoic sediments is dissected by the 100–270 m high water divides of the Kuloi Plateau and the Pokshenga Highland, whereas the Kanin Ridge is composed of Late Precambrian metasediments 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 to 80 m a.s.l. Deep troughs in the White Sea lying more than 50 m below the 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 m thick, except for uplifted regions where strongly weathered bedrock and boulder 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 has been provided by Demidov et al. (2004). Ramsay (1911) recognized three till units...
separated by shell-bearing sediments, i.e. 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 centres of glaciation. Two or three till beds have since been recognized in the Kanin Peninsula area (Kalayanov & Androsova 1933; Lyutkevich 1947; Spiridonov & Jakovleva 1961). Spiridonov & Jakovleva (1961) also supported the idea that the upper till was of Scandinavian origin. This is confirmed by Kjær et al. (2006) and by Larsen et al. (2006b). The presence of Scandinavian boulders, young eskers and sandy kames in the western parts of the 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 (LGM) (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 photographic interpretation. Edemsky (1931), Devyatova (1982) and Filipov & Borodai (1987) suggested that the middle parts of the Pinega and

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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).
Mezen rivers remained unglaciated after the Eemian (Mikulinian) interglacial, while others argued that SIS or the Kara Sea Ice Sheet reached the area either during the Early Weichselian (Krasnov 1971) or at the end of the Weichselian (Aseev 1974; Lavrov 1991; 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 and satellite images of scale 1:500 000 and 1:1000 000 (Zatulskaya 1984a, b, c). Field observations in selected areas served as ground control. In Fig. 2 we have distinguished between end moraines, hummocky moraine, kames, sandar, glacial lake basins, bedrock or weathered rock, bedrock cliff, valley trains and plateaus.

River and coastal sections were logged and facies associations were interpreted in terms of depositional environment. Diamictites were described, classified and interpreted following common guidelines (cf. Krüger & Kjær 1999; Kjær et al. 2001). Directional properties of till units include striations on subtilt bedrock and boulder pavements and, most frequently, data from clast fabric analyses. Samples of 25 clasts were obtained from a subhorizontal surface c. 25 × 25 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 (V1) and eigenvalues (S1 and S3) for each sample were computed using the program SpheriStat, version 2.0 (Pangaea Scientific) based on Mark (1973, 1974). According to Anderson & Stephens (1971), clasts are considered to have statistical preferred orientation if the S1 and S3 eigenvalues are ≥0.52 and ≤0.17, respectively. In order to verify the significance of eigenvectors and test for bimodal distribution of clasts, contoured diagrams were prepared following the approach by Kamb (1959).

When unit boundaries comprise angular unconformities indicating glaciectectonic 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; in particular, Kola Peninsula nepheline syenites are found dispersed in the northern part of the 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, Riso, Denmark and the results are listed in Table 1. Doses were measured 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 inference that the sediments were frozen during most of the Weichselian; in periods of thaw they would have remained in the saturated groundwater zone until sub-recent erosion by modern rivers lowered the local groundwater 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 do 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%. This mean change is small compared to the typical overall age uncertainty of ~10%, as is the standard deviation of this change. All uncertainties resulting from these possible variations in burial depths are <55% of the total uncertainty in the ages, and the ages critical of the dating of the LGM (those between 30 and 12 kyr) are changed by <1000 years. In summary, the possible 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 &

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

<table>
<thead>
<tr>
<th>No.</th>
<th>Fig.</th>
<th>Strat. Pos.</th>
<th>Locality</th>
<th>Risø code</th>
<th>Sample no.</th>
<th>Age (kyr)</th>
<th>Dated material</th>
<th>Dose rate</th>
<th>W.C.</th>
</tr>
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<td>Below</td>
<td>Tarkhanov river</td>
<td>001020</td>
<td>00421</td>
<td>80±6</td>
<td>Glaciolacustrine</td>
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<td>00423</td>
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<td>01508</td>
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<td>00406</td>
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<td>2.26</td>
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<td>00441</td>
<td>19.1±1.5</td>
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<tr>
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<td>00442</td>
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<td>99422</td>
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<td>00437</td>
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<td>00439</td>
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<td>00440</td>
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<td>00441</td>
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<td>Marine sand</td>
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<td>1.72</td>
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<td></td>
</tr>
</tbody>
</table>

Cosmogenic exposure dates

| 1   | Below | Tarkhanov river | 00418 | 51.5±3 | Quarts vein |
| 2   | Below | Tarkhanov river | 00419 | 55.0±4 | Quarts vein |
| 3   | Above | Tarkhanov river | 00420 | 18.0±1.6 | Quarts vein |

Fig 3. Stratigraphic logs from selected sections along the northern shores of the Kuloi Plateau.
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 Fig. 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 has previously been 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 Vologda 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 1969). Previous detailed stratigraphical 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 and 21 kyr (Table 1). Scandinavian till unconformably overlies these sediments and may have been glaciotectonically deformed. The basal part of the till often shows sheared and folded lenses of Quaternary deposits. Subtill and intertill glaciodynamic structures, striations on clasts and clast fabric orientation indicate glacier movements from westerly and south-westerly directions (Fig. 4, sites 8, 9, 11). Occasionally, ice-dammed lake sediments of similar stratigraphic position contain dropstones of Scandinavian origin (Fig. 3, sites 8–13, 15, 49). The till ranges in thickness from 2 to 6 m 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 Scandinavian till or they constitute its lateral equivalent. Deposits also show ice-wedge casts and other periglacial structures overlain by fluvial deposits containing plant detritus. These sediments yield ages of 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 westward, we assume the limit of SIS to be located near the east bank of the mouth of the Kuloi River (Figs 1, 3). Even though the topography is rather flat and glacial landforms are blurred by peat bogs, an 8-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 Figs 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 optically stimulated luminescence (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 at the western and northern Kanin Peninsula.

At the Tarkhanov cliff section (Fig. 1, site 1) a 40 to 45-m-high terrace of Weichselian sediments is banked upon the Precambrian metasediments 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 to 15-m-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, 80 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-m-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 the quartz veins on the bedrock surface at different altitudes
indicates ages of exposure of 18.6 kyr below c. 50 m a.s.l., whereas at adjacent inland altitudes around 100–115 m a.s.l. ages indicate exposure of around 51 and 56 kyr (Linge et al. 2006). OSL dates of sediments in the kame hills at altitudes above 130 m on the inland plateau a few kilometres northward gave OSL ages of 71 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 m 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 the Scandinavian Ice Sheet, the flow direction of which was either deflected 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 suggest that the last SIS reached an altitude up to 60–70 m a.s.l. and moved along the White Sea slope of the Kanin Ridge before 18 kyr ago. Because the top of the Kanin Ridge reaches altitudes of 100–180 m 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. However, new data from sections at Ezhuga, Karpogory and Verkola (Fig. 1, sites 19, 20, 22 and Fig. 7) suggest that the limit should 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 m. Sediments with Eemian marine shells lacking a till cover are widespread in surface exposures (Filippov & 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

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**Fig. 4.** Clast fabric diagrams from selected sections stereographically plotted on the lower hemisphere.
The age of the glaciofluvial and glaciolacustrine sediments below and above (Fig. 7) suggest that the till has been deposited by an ice sheet from north–northwest in the Early Weichselian or possibly in the Late Saalian. On the left side of the Ezhuga River, marine interglacial sediments resting on the Moscowian (Late Saalian) Scandinavian till are, in turn, covered by a 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; we propose that the limit is possibly marked by the many end-moraine ridges and meltwater channels that occupy the northwestern part of the area (Zatulskaya 1984a, b). Glaciolacustrine sand at Ezhuga (site 19), with an age of 26.6 ± 1.9 kyr and lying above clast-bearing mud (Fig. 7, Table 1), does not provide conclusive evidence as to whether the last SIS did override this area. West of the end-moraine belt, i.e. east of 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 suggest that the limit of the last SIS in this area is correct compared to previous estimates (Figs 1, 2, 7), but 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 re-deposited marine shells of Eemian age.

Position of the last SIS in the Arkhangelsk region

Since Scandinavian till is clearly absent in sections along the coastal area of the western Kanin Peninsula, the ice margin continued from Cape Kanin Nos probably across the eastern White Sea floor southwest to Cape Kargovsky (Figs 1, 2). Cape Kargovsky is the easternmost 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 water divide between the Kuloi and Mezen Rivers and crossed the valley of the Kuloi River near the mouth of the Soyana River. The ice front turned westward from the mouth of the Soyana River and touched the steep, 50 to 80-m-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 that guided meltwater towards the east (Krasnov 1971; Legkova & Schukin 1972).

During the maximum extent, the 180-m-high southeastern part of the Kuloi Plateau was ice free (Edemsky 1931); we propose that the ice margin only rimmed the plateau. From the south, an ice tongue moved eastward along the Pinega River lowland to altitudes of 30–60 m 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 of Shilega and Karpogory, where Eemian marine sediments with whale bones are not covered by till (Devyatova 1982). The ice margin continued along a discontinuous chain of hills and ridges south–westward to join terminal formations which truncate the Pokshenga Hills from the north and west between sites 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 has been supported for similar reasons by Larsen et al. (1999) and Atlasov et al. (1978).

In our present configuration of the last SIS, the data suggest a smaller 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 of the Karpogory village in the 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, 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 water divide. It was proposed to mark an end-moraine zone (Zatulskaya 1984c; Lavrov 1991) and according to Demidov et al. (2004) it depicted 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 merely reflects 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 indicates ice advance, hence the distribution of a till sheet or glaciotectonic deformation can be used to reconstruct the extent of the ice sheet. In our opinion, it is likely that wet-based ice lobes extended eastward during the LGM ice-sheet build-up. However, frozen bed conditions could have been regained in the marginal zone when ice lobes reached higher and permafrozen ground east of the White Sea basin, as seems to be the case with the ice cover on the high grounds of the Kola Peninsula (Hättestrand 2006). Although we are unable to exclude this possibility, there is compelling 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 ice extent (Fig. 9). Subtill dates range from 32 to 15.9 ± 1.2 kyr, whereas dates from above the Bobrovo till range from 17.2 ± 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 and 16 kyr ± 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 the maximum extent was most likely 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 have not added significantly to previous estimates based on data restricted to the Severnaya Dvina area (Larsen et al. 1999). In addition, exposure dating on the Kanin suggests that SIS began to retreat from the northwestern part of Peninsula as early as 18.6 kyr ago, and therefore neither contradicts this age estimate nor provides a better accuracy. An average age of the maximum extent of the eastern part of the Scandinavian Ice Sheet placed between 18 and 16 kyr is at least 3–5 kyr and perhaps as much as 7–9 kyr.
younger than previously proposed by Arslanov et al. (1970, 1971).

Deglaciation drainage pattern

Palaeogeographical reconstructions depict the position of the retreating ice margin, the distribution of ice-dammed lake basins and the drainage directions of meltwater from the time of the largest extent of SIS (Fig. 8) until the Allerød interstadial (Fig. 10A–D). Ice-dammed lake and river terrace 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 5 kyr ago 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 the 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 m a.s.l. and was controlled by a threshold located on the lowermost part of the Kanin Peninsula on the water divide 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 m of fluviatile sediments. The valley continues as a submarine canyon eastward at depths around 15–30 m below sea level. It is more than 20 km length, up to 1 km in width and up to 30 m in depth. The ice-dammed lake discharged to the Chyosha Bay. Laminated clay on the bottom of this lake is located at altitudes of 5–17 m a.s.l. (Kjær et al. 2003) and can be seen in the section from sites 13 and 15 (Figs 1, 3). The elevated river terraces observed by Zekkel (1939), with altitudes of 5–6, 8–10, 20 m a.s.l. on the lower Mezen and Pyoza rivers, may represent falling water level in the lake with time. A terrace at 15 m on the Kanin Peninsula may correspond to one of these stages.

Kuloi ice lake

The ice-dammed Lake Kuloi 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 into 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 no more than 15–20 m 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 are composed of coarse sand and have an altitude of 5–20 m 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 m a.s.l. with a gradual drop to 16 m. 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 m a.s.l. and was controlled by a threshold of 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 m a.s.l., are known from both 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 m 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 ± 1.6 kyr ago, but the Pinega River maintained its discharge, now supplied with meltwater from the Severnaya Dvina ice lake. As the ice melted further away, discharge carved a deep spillway at altitudes of 18–20 m a.s.l. on the Kuloi River–Pinega River watershed (Fig. 10D). The ages of events are based on dates of aeolian sediments lying at altitudes about 30 m a.s.l. in the former lake basin at the Kulogora village on the Pinega–Kuloi water divide (Figs 1, 7, site 18).

**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 m a.s.l. (Atlasov et al. 1978; Arslanov et al. 1984) and runoff 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 m a.s.l.; discharge was redirected into the Mezen Bay through the middle part of the 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 m a.s.l. and a new 50-m threshold developed on the water divide of the Pokshenga–Pukshenga rivers, while runoff was directed downstream 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 m a.s.l.; discharge was redirected into the Mezen Bay through the middle part of the Pinega and Mezen rivers (Fig. 10A).

By the time the active ice margin retreated from the mouth of the Pinega River, the level dropped from 50 to 18 m...
Fig. 10. Palaeogeographical reconstructions picturing four stages (A–D) of Lateglacial ice retreat, deglaciation and development of ice-dammed lakes and run-off patterns in northwest Russia.
a.s.l., 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 18–20 m 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 the Psaryovo section (Fig. 1, site 41) where plant remains with an age 13.2 kyr occur in fluvial sand at 25 m 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 the Psaryovo section (Fig. 1, site 41) ranging from altitudes of 34 m 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 m above the present river level (Fig. 11). The sediments are exposed in the river banks of the present 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 beneath (Lyså et al. 2001).

Overlying basal till with Scandinavian provenance components of four major depositional events can be 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. At some sites, imbricate gravel or cross-beded sand with an erosional lower contact suggests deposition in meltwater streams, and this possibly signals the lowering of the base level caused by a drop in the 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 meltwater 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 of freshwater molluscs into the lakes. The flora includes species such as Betula, Salix, Populus, Larix and Picea (Lysá et al. 2001).

Finally, deposition of rhythmically laminated mud with occasional diamict lenses, dropstones and beds of sand was followed by an episode of erosion caused by a 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 inundated the White Sea around 11 kyr ago (Koshechkin et al. 1977) and subsequent regression began at the onset of the Boreal period about 9100 years ago (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 Fig. 12. The SIS invaded southern and central Finland less the 25 kyr ago, as indicated by dating of mammoth tusks pre-deposition dating of till by the last SIS (Ukkonen et al. 1999; Lunkka et al. 2001). In the following 5–9 kyr, the glacier front had advanced at an average velocity 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 the velocity. Corridors of rapidly flowing ice separated by slower flowing ice constituted ice-divide zones.

Our results suggest that glacier tongues and lobes, probably less than 300 m 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 still has to be thoroughly tested, is based on theoretical models of similar dynamic behaviour of terrestrially based ice streams as predicted by Stokes & 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 prerequisites, including little or no topographic control, and special bed lithologies for the location of potentially 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 territorially 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 laterally confined ice flow occurred around the whole SIS, factors such as the presence of water-filled basins and deformable substratum must have played a major role. From the northern North Sea to the circum-Baltic region, rapidly flowing ice streams bounded by slow flowing inter-stream areas dominate the pattern of Late Weichselian glaciation. With the exception of the main advance inter-stream event in Denmark (22–19 kyr BP), the Norwegian Channel Ice Stream (26 kyr BP; Sejrup et al. 1994), the Kattegat Ice Stream (29–25 kyr BP) and the Young Baltic Ice Streams (19–15 kyr BP), all advanced from SIS across marine waters or ice-dammed lakes. These advances show flow patterns and left behind morphological features similar to those of terrestrial-based ice streams (Boulton et al. 2001; Clark et al. 2003 and references therein; 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 the main ice sheet. Moreover, glacier flow velocities were probably reduced and the ice considerably thinner in areas of elevated Pre-Quaternary bedrock. In these inter-lobe areas, third-order flow divides sensu Boulton
et al. (2001) could have been located. Along the northeastern part of the ice sheet in Russia, the Barents Sea–Kanin fast-flowing ice moved from the west and northwest 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 the 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, 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 m a.s.l. The Dvina and Pinega ice tongues (Fig. 12, IIB, IIC) flowed from the northwest 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 m a.s.l. (Aseev 1974).

Decay of the ice sheet

Event-stratigraphic models in the region often relate Lateglacial climatic variations to the ice-marginal positions and recessive stages of the SIS (Krasnov 1971; Aseev 1974), but our data do not challenge 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

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 Bolling–Allerød ice marginal position. Data from the present study are compiled with data from Aseev (1974), Gey & Malakhovsky (1998), Ekman & Iljin (1995), Lunkka et al. (2001) and Hättestrand (2006).
caused the development of kettle holes with 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/C1 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 rapidly flowing ice lobes are overprinted by landforms that relate to aerial downwasting (cf. 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 circled the highland of the Kola Peninsula. Between the maximum position of SIS and the Neva stage termini, a several hundred thousand km² large zone of downwasting contains a complex dead ice relief, including fields of kame-like hills and hummocky moraine. Over large areas glaciolacustrine sediments may blur the original shape and size of the dead ice landforms.

We assume that long and thin lobes of fast-flowing ice became 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 ago and lasted perhaps until the beginning of the Allerød interstadial (Fig. 13). Bolling sediments rarely occur in this zone probably because the larger part of the area was covered by unstable dead ice. The age of the oldest organic remnants in sediments from lakes across the downwasting zone belongs to the Allerød interstadial (Ekman & Ilijin 1995; Davydova et al. 1998). Deposition of organic debris in small lakes on the watersheds covered by dead ice began as late as the Preboreal and Boreal (Demidov & Lavrova 2001; Wohlfarth et al. 2002, 2004). Huge areas in the peripheral parts of the SIS therefore 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 thick overburden retarded heat transfer, thereby delaying downwasting.

From the Allerød interstadial and onwards, decay of the active ice margin changed to frontal-type deglaciation. The ice sheet rested on crystalline bedrock and was not enriched with debris to the same degree 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. 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 & Ilijin 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 some time between 20 and 15 kyr ago; in the Arkhangelsk region this probably occurred around 18–16 kyr ago. This position was reached up to 10 kyr after the maximum extent in the southwestern part of SIS.
- At the maximum, the ice margin stretched from the northwest part of the Kanin Peninsula across the 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.
- In the zone between the former margin of SIS and the Neva terminal belt, aerial downwasting caused the inversion of landscapes. Deglaciation sediments are composed of gravity flows of diamicton 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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References

Anderson, T. W. & Stephensen, M. A. 1971: Tests for randomness of

Bergen) and Clas Hättestrand (Stockholm University), who put
carrying out the luminescence dating and readily discussed the results.

Demidov, I. N., Houmark-Nielsen, M., Kjær, K. H., Larsen, E.,


glacial maxima in the Arkhangelsk region of Northwest Russia.

In Ehlers, J. & Gibbard, P. L. (eds.): Quaternary Glaciations – Extent


Devyatova, E. I. 1969: Deglaciation of Valdai glaciation and

lateglacial history of Baltic and White Sea. In Gerasimov, I. P.

(ed.): Poslednii ledukonnyi pokrov na severo-zapade evropskoi


Devyatova, E. I. 1982: Late Pleistocene environments as related to

human migrations in the Severnaya Dvina basin and in Karelia. 156


Edensky, M. B. 1931: Geological investigation in the basins of the

Rivers Pina, Kuloi and Mezen at 1929. Trudy geologicheskogo


Ekmann, I. & Iljin, V. 1995: Deglaciation, the Younger Dryas end

moraines and their correlation in Russian Karelia and adjacent

areas. In Ehlers, J. & Gibbard, P. L. (eds.): Glacial Deposits in


Filippov, V. V. & Borodai, L. V. 1987: Key sections of the Mikulino

marine deposits in the north of the Mezen hollow. Vestnik

Leningradskogo Gosudarstvennogo Univeristeta, Geologiya, Geogra-

fii 21, 72–74 (in Russian).

Funder, S., Demidov, I. & Yelovicheva, Ya. 2002: Hydrography and

mollusc faunas of the Baltic and White Sea – North Sea seaway in

the Eemian. Palaeogeography, Palaeoclimatology, Palaeoecology

184, 275–304.

Ganeshin, G. S., Krasnov, I. L., Marinov, N. A. & Murzaeva, V. E.

(eds.) 1980: Map of Quaternary Deposits of Eurasia. Scale

1:5000000. Ministerstvo geologii SSSR, VN II zarubezhgeologiya

(in Russian).

Geology of USSR 1963: Arkhangel'sk and Vologda Districts and Komi


Gey, V. P. & Malakhovskiy, D. B. 1998: On the age and expansion of

maximal late Pleistocene glacial thrust in western part of the

Vologda area. Izvestia Rossiiskogo geograficheskogo obozhestv 1,

43–53 (in Russian).

Hättestrand, C. 2006: The glacial morphology of Kola Peninsula and

adjacent areas in Murmansk Region, Russia. Journal of Maps

v2006, 30–42.

Houmark-Nielsen, M., Demidov, I., Funder, S., Grossfeld, K., Kjær,


and Middle Valdai glaciations, ice-dammed lakes and periglacial

interstadials in northwest Russia: new evidence from the


Houmark-Nielsen, M. & Kjær, K. H. 2003: Southwest Scandinavia,

40–15 kyr: palaeogeography and environmental change. Journal of

Quaternary Science 18, 769–786.

Kalyanov, V. P. & Androsova, V. P. 1933: Geographical observation


Kamb, W. B. 1959: Ice petrofabric observations from Blue Glacier,

Washington, in relation to theory and experiment. Journal of

Geophysical Research 64, 1891–1909.

Kitagawa, H. & van der Plicht, J. 1998: A 40,000 year varve

chronology from lake Sietus, Japan: extension of the

$^{14}$C calibration curve. Radiocarbon 40, 515–530.

Kjær, K. H., Demidov, I., Houmark-Nielsen, M. & Larsen, E. 2001:

Discrimination between easterly- and westerly-flowing Valdai ice

streams in Arkhangelsk region, Northwest Russia. Global and

Planetary Change 31, 201–214.

Kjær, K. H., Demidov, I. N., Larsen, E., Murray, A. & Nielsen, J. K.

2003: Mezen Bay – a key area for understanding Weichselian

glaciations in northern Russia. Journal of Quaternary Science

18, 73–93.

Kjær, K. H., Larsen, E., Funder, S., Demidov, I., Jensen, M.,

Håkansson, L. & Murray, A. 2006: Eurasian ice sheet interaction in

northwestern Russia throughout the late Quaternary. Boreas 35,

000–000 (this issue).
orie, Onega Bay of White Sea. Boreas 28, 386–402.

Kvasov, D. D. 1975: Late Quaternary History of Large Lakes and 

Lagerlund, E. 1987: An alternative Weichselian glaciation model, 
with special reference to the glacial history of Skåne, south-

Larsen, E., Lyså, A., Demidov, I., Funder, S., Hounam-Nielsen, M. & 
Kjær, K. H. 1999: Age and extent of the Scandian Ice Sheet 

Larsen, E., Kjær, K. H., Demidov, I., Funder, S., Gressfeldt, K., 
Hounam-Nielsen, M., Jensen, M., Linge, H. & Lyså, A. 2006a: 
Late Pleistocene glacial and lake history of northwestern Russia. 
Boreas 35, 000–000 (this issue).

Larsen, E., Kjær, K. H., Jensen, M., Demidov, I., Hakkanson, L. & 
Pauš, Aa. 2006b: Early Weichselian palaeo-environments re-
constructed from a mega-scale thrust-fault complex, Kanin Peninsula, 
northwestern Russia. Boreas 35, 000–000 (this issue).

Lavrov, A. S. 1991: Map of the Quaternary Deposits. Sheet Mezen 
Scale 1:1000000. VSEGEI, Leningrad (in Russian).


Legkova, V. G. & Schukin, L. A. 1972: Belts of terminal formations 
in western part of Arkhangelsk district. In Goretysy, G. I., 
Pogulyaev, D. I. & Shik, S. M. (eds.): Kraevye obrazovania 

Linge, H., Larsen, E., Kjær, K. H., Demidov, I., Brook, E. J., 
Raisbeck, G. M. & Yiou, F. 2006: Cosmogenic 10Be exposure 
dating across Early to Late Weichselian ice-marginal zones in 
northern Russia. Boreas 35, 000–000 (this issue).

Lunichka, J. P., Saarnisto, M., Gey, V., Demidov, I. & Kiseleva, V. 
2001: The area and timing of the Last Glacial maximum in the 
Valdaian (Weichselian) cold stage in Vologda and adjacent areas of 

Lyså, A., Demidov, I., Houmark-Nielsen, M. & Larsen, E. 2001: Late 
Pleistocene stratigraphy and sedimentary environment of the 
Arkhangelsk area, northwest Russia. Global and Planetary Change 
31, 179–199.

Lyutkevitich, E. M. 1947: The history of Kanin peninsula in 
Quaternary time. Izvestia VGO 79, 8–73 (in Russian).

Mark, D. M. 1973: Analysis of axial orientation data including till 

101–104.

Murray, A. S., Marten, R., Johnston, A. & Martin, P. 1987: Analysis 
of naturally occurring radionuclides at environmental concentra-
tions by gamma spectrometry. Journal of Radioanalytical and 

Murray, A. S. & Wintle, A. G. 2000: Luminescence dating of quartz 
using an improved single-aliquot regenerative-dose protocol. 
Radiation Measurements 32, 57–53.

Pleshitseva, E. S. 1977: Changes of paleogeography conditions in 
North-Dvina lowland at late-postglacial time. Priroda i hayzatsyo 
Severo 6, 39–47 (in Russian).
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