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Ice-dammed lakes and rerouting of the drainage of northern Eurasia during the Last Glaciation

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Abstract

During the Quaternary period, ice sheets centred over the Barents and Kara seas expanded several times onto mainland Russia and blocked northflowing rivers, such as the Yenissei, Ob, Pechora and Mezen. Large ice-dammed lakes with reversed outlets, e.g. toward the Caspian Sea, formed south of these ice sheets. Some lakes are reconstructed from shorelines and lacustrine sediments, others mainly from ice-sheet configuration. Ice-dammed lakes, considerably larger than any lake on Earth today, are reconstructed for the periods 90–80 and 60–50 ka. The ages are based on numerous optically stimulated luminescence (OSL) dates. During the global Last Glacial Maximum (LGM, about 20 ka) the Barents–Kara Ice Sheet was too small to block these eastern rivers, although in contrast to the 90–80 and 60–50 ka maxima, the Scandinavian Ice Sheet grew large enough to divert rivers and meltwater across the drainage divide from the Baltic Basin to the River Volga, and that way to the Caspian Sea. Climate modelling shows that the lakes caused lower summer temperatures on the continent and on the lower parts of the ice sheet. The final drainage of the best mapped lake is modelled, and it is concluded that it probably emptied within few months. We predict that this catastrophic outburst had considerable impact on sea-ice formation in the Arctic Ocean and on the climate of a much larger area.

1. Introduction

Ice sheets centred over the Barents and Kara seas moved up-slope when they expanded onto mainland Russia (including Siberia). Therefore the north-flowing rivers were diverted and ice-dammed lakes formed in front of the ice sheets. In a similar way the Scandinavian Ice Sheet dammed lakes and diverted rivers in western Russia. These concepts have been understood for decades (Kvasov, 1979; Grosswald, 1980), but the impact and timing of the blocking events have been uncertain.

Rerouting of the north-flowing rivers must have occurred during all large Quaternary glaciations that affected northern Russia. However, the glaciation pattern, and thus the diversion of rivers, certainly changed from one glaciation to another. In this paper we analyse the drainage history during the Last

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Glaciation, named Zyryanka in Western Siberia, Valdai in European Russia, and Weichsel in Western Europe. Our presentation is structured chronologically. We use the western European chronostratigraphic terminology, i.e. the Weichselian, considered as nearly synchronous with marine isotope stages (MIS) 5d-2, and subdivided into the Early Weichselian (MIS 5d-5a), the Middle Weichselian (MIS 4-3), and the Late Weichselian (MIS 2) (Mangerud, 1989). Rerouting of the drainage occurred in five areas, although with interplay between some of them (Fig. 1): (1) the Taimyr Peninsula, (2) the West Siberian Plain, (3) the Pechora Lowland, (4) the White Sea Basin, and (5) the area draining toward the Baltic Sea.

Our analysis of the drainage history is mainly based on the results of the European Science Foundation project "Quaternary Environment of the Eurasian North" (QUEEN). We aim at an overview and for observations will mainly refer to the relevant publications. However, some previously unpublished observations will be documented in more detail. This work is a companion paper to the synthesis of the ice sheet history (Svendsen et al., 2004) and we will therefore also discuss potential drainage rerouting caused by glaciations for which the drainage history is not fully documented. The main reconstructions are given in three maps (Figs. 2–4).

2. Methods

The "QUEEN-teams" investigating different areas have, to different degrees, employed field observations, photogeological studies, and dating programs; details are given in the cited papers. On the West Siberian Plain, only the Lower Ob area was investigated as part of the QUEEN efforts; the rest of this huge region has been assessed only from data in previous Russian works.

The chronology is mainly based on optical stimulated luminescence (OSL) dates obtained at the Nordic Luminescence Laboratory in Risø, Denmark. All dates are performed with the SAR protocol (Murray and Wintle, 2000), although some of the older ones also include measurements undertaken using the multiple aliquot SARA (Mejdahl and Bøtter-Jensen, 1994) and additive dose protocols. Until about 1999 the dose rate term was measured using field gamma spectrometry and/or laboratory beta counting; thereafter the dose rate was determined using laboratory gamma spectrometry (Murray et al., 1987). These improvements in technology provide improvements in calculated accuracy and precision in the more recent dates (post 1999), and probably make them more reliable. Nevertheless, some

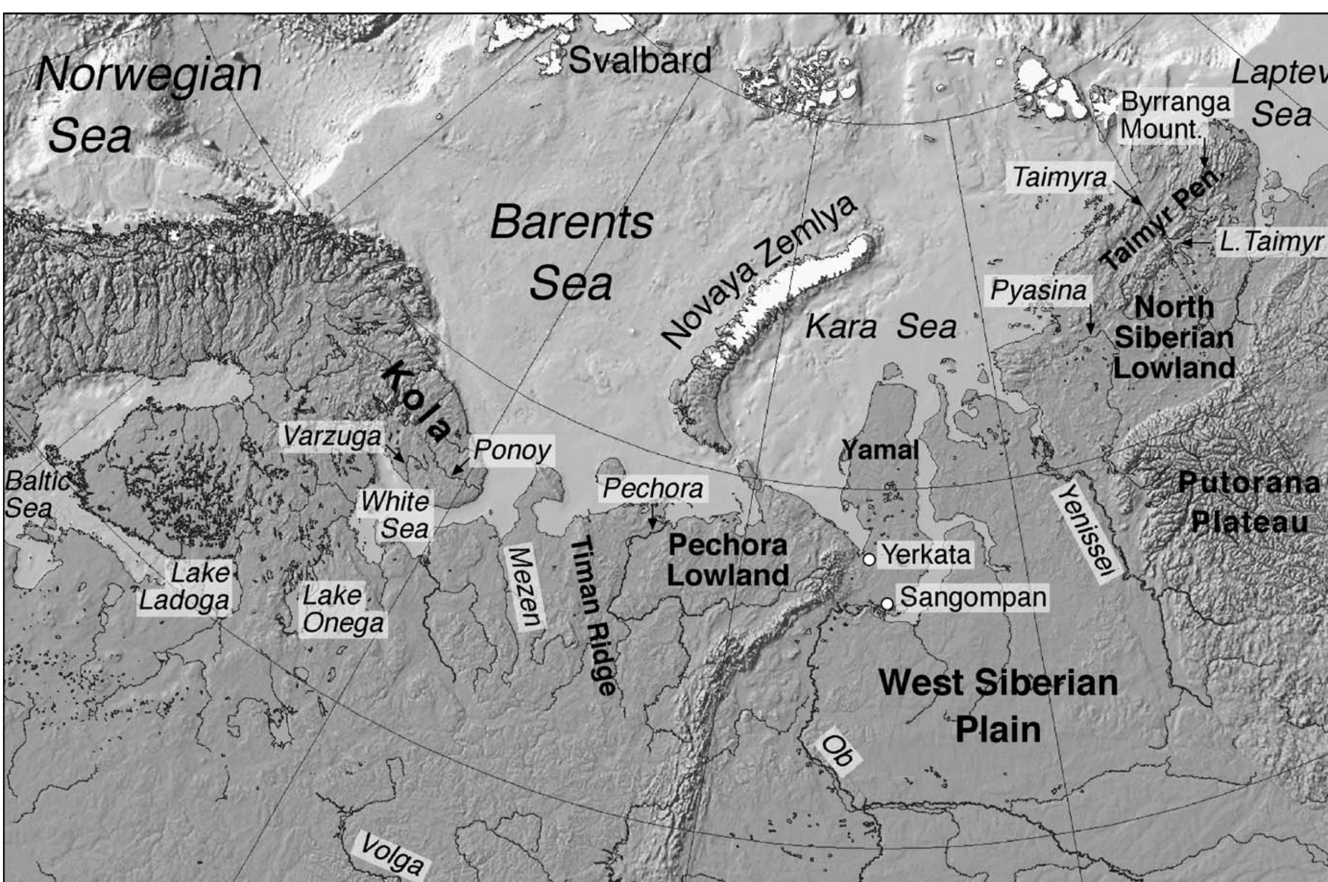

Fig. 1. Key map with names used in the text.

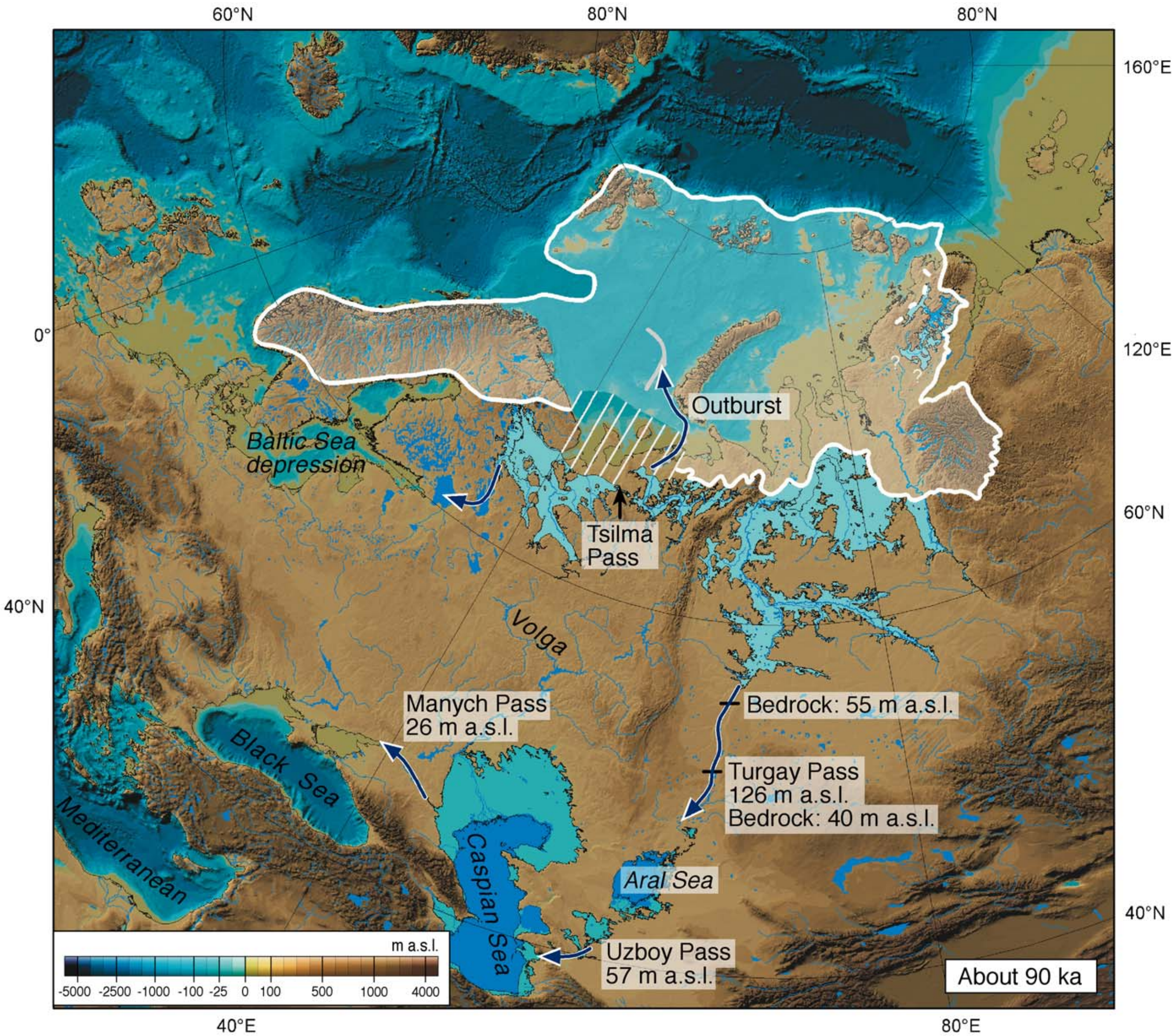

Fig. 2. Reconstruction of ice-dammed lakes and rerouting of rivers during the Early Weichselian, about 90–80 ka. Ice margins are taken from Svendsen et al. (2004). In the hatched area the ice margin position is unknown, probably because it was overrun by the 60 ka ice advance. Stippled line on Taimyr shows a retreat phase damming a lake. Blue arrows show outlets. The arrow in the Barents Sea shows the longest modelled outburst route for Lake Komi, and the corresponding western ice margin. The shorter and more probable routes have the same starting point. See text for discussion. Sea level is lowered 50 m (Chappell et al., 1996) without considering any isostatic depression.

assumptions remain, such as the stability of radionuclide activities, and thus of dose rates.

The area and volume of all the larger reconstructed lakes are estimated in the same way as in Mangerud et al. (2001a), i.e. using the Global Land One-km Base Elevation (GLOBE) model (GLOBE-Task-Team, and-others, 1999), combined with the International Bathymetric Chart of the Arctic Ocean (IBCAO) bathymetric grid model (Jakobsson et al., 2000). We have used the ice margins as shown on the maps (Figs. 2–4). These are taken from Svendsen et al. (2004), except that in the Pechora Lowland we have used the Harbei moraines for 90–80 ka and omitted the Laya-Adzva and Rogovaya ridges because Lake Komi extended to the Harbei moraines (Fig. 5).

We have filled up the lakes to assumed surface level, and calculated the areas and volumes using the present-day surface topography. For Lake Komi and the ice-dammed lakes on Taimyr, mapped shorelines are at higher altitudes closer to the ice-damming margin, indicating a glacioisostatic tilting. This information has been used for the areal and volume reconstructions of Lake Komi and the lakes on Taimyr, whereas for all the other lakes a horizontal lake surface have been used.

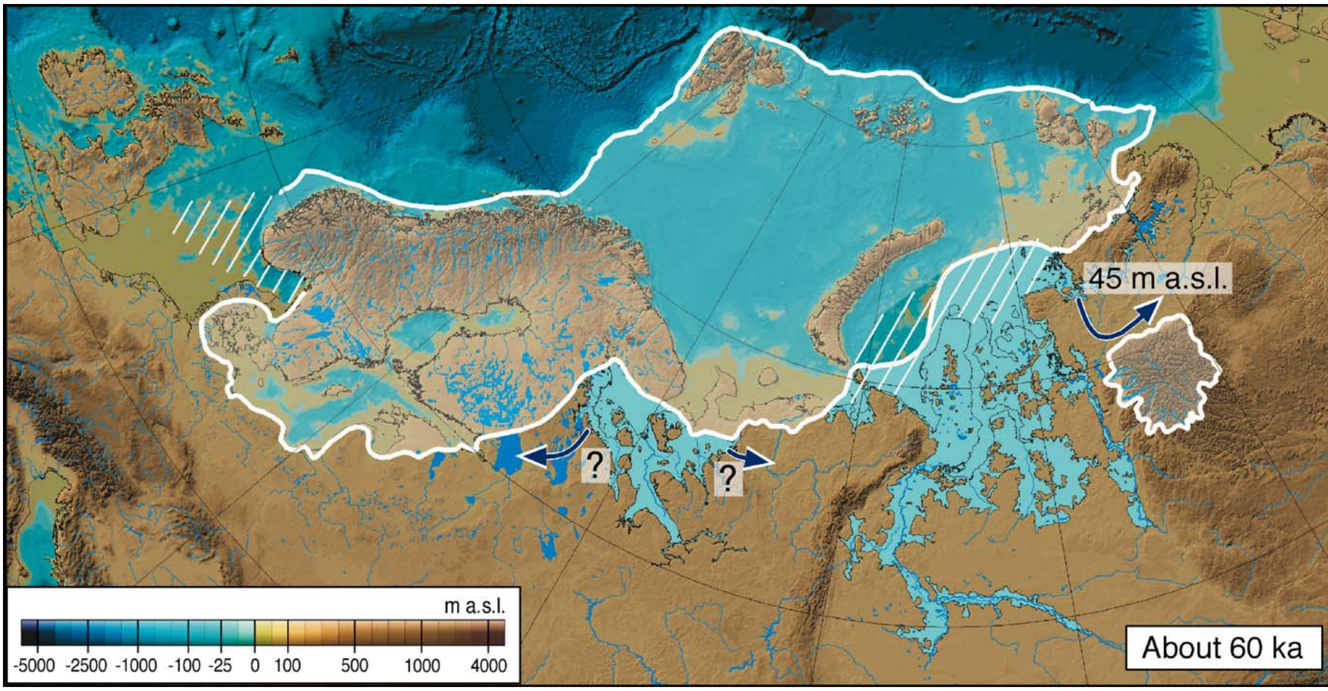

Fig. 3. Reconstruction of ice-dammed lakes and rerouting of rivers during the Middle Weichselian, about 60–50 ka. Ice margins are taken from Svendsen et al. (2004). In the hatched area, position of ice margin is unknown. The lakes are to a large extent predictions based on the configuration of the ice margin. Thus they represent visualisations of hypotheses that should be tested. Arrows show assumed outlet paths. Sea level is lowered 60 m (Chappell, 2002).

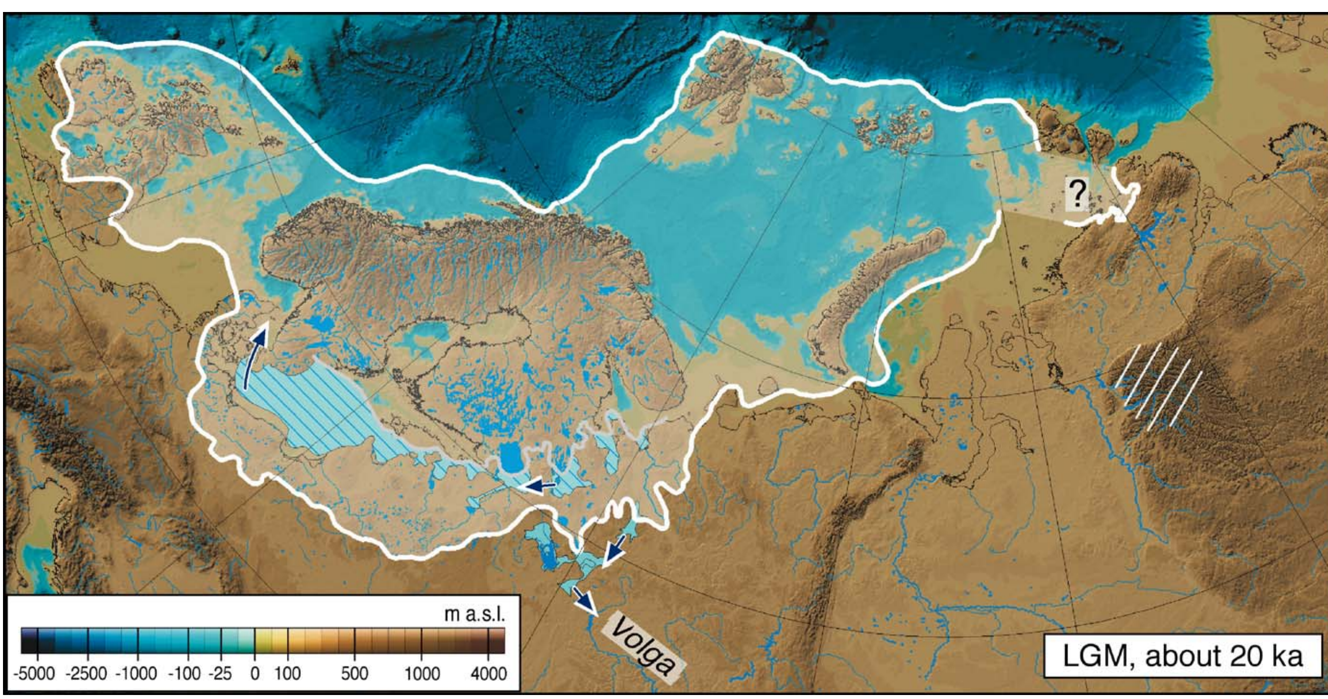

Fig. 4. Reconstruction of ice-dammed lakes and rerouting of rivers during the LGM about 20 ka. Inside the LGM limit is also shown younger (about 14 ka) ice-dammed lakes around Lake Onega and in the Baltic Sea depression, and their outlets. The misfit between the latter lake and the present Baltic Sea coast is due to subsequent glacioisostatic tilting. Sea level is lowered 120 m everywhere.

3. Early Weichselian lakes and rerouting of rivers

The first Weichselian ice advance of the Barents–Kara Ice Sheet that is documented in northern Russia (Svendsen et al., 2004) dammed the lakes in Fig. 2. Most dates are obtained from shorelines of Lake Komi in the Pechora Lowland, where 29 OSL dates indicate ages of 90–80 ka. We therefore describe this paleolake first.

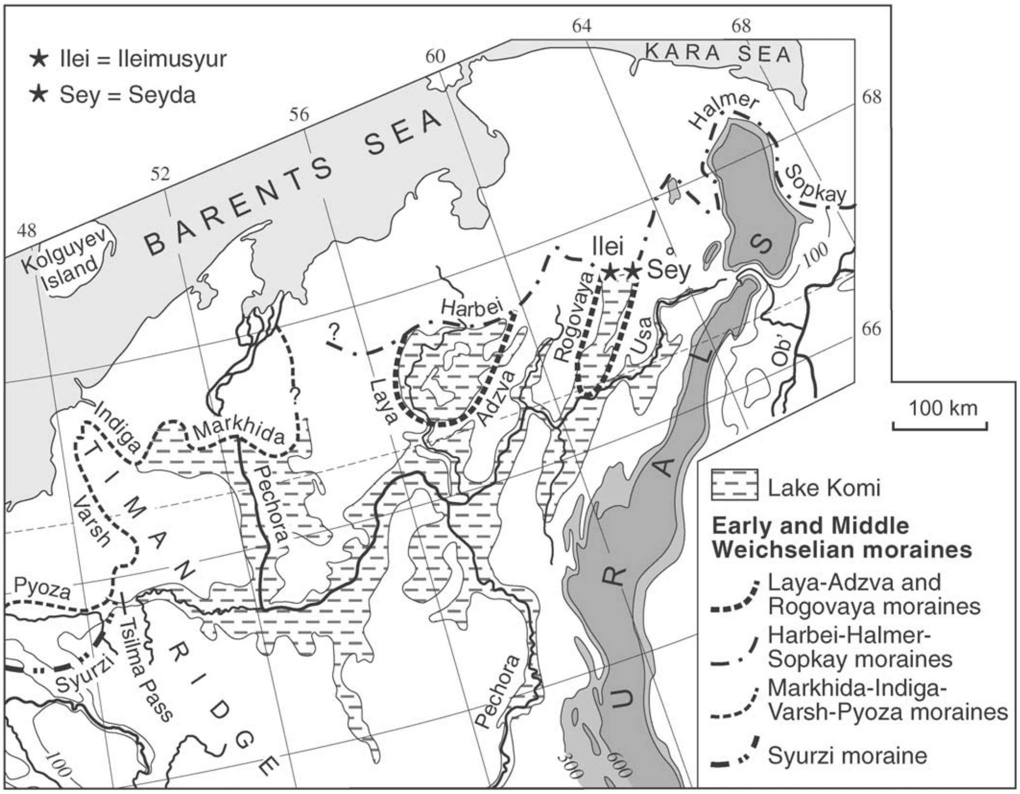

Fig. 5. Map of the Pechora Lowland. The suggested extension of Lake Komi through the Tsilma Pass into the White Sea Basin is also shown. The Harbei–Halmer–Sopkay moraines are considered to represent the ice margin that dammed Lake Komi about 90–80 ka. Their continuation toward west is unknown and we have limited the lake there along the younger Markhida–Indiga moraines.

3.1. Lake Komi in the Pechora lowland

The Pechora Lowland is bordered by the Urals in the east, the Timan Ridge in the west, and in the south by the continental water divide between the Barents Sea and the Volga River/Caspian Sea (Fig. 1). Strandlines of the former ice-dammed Lake Komi have been mapped for a distance of more than 2000 km (Fig. 2), using air photos and satellite images (Fig. 6) (Astakhov et al., 1999; Nikolskaya et al., 2002). These morphologically expressed shorelines are discontinuous and account for only 25% of the perimeter of the lake, yet it is surprising that so much of a 90–80 kyr old shoreline is preserved. The shoreline elevation ranges from 90 m a.s.l. in the south to 110 m in the north, reflecting a larger glacio-isostatic depression near the ice margin in the north.

There are two main types of deposits reflecting the Lake Komi shorelines. The first is from wave dominated shores, including some river deltas. This type makes up the long strandlines mapped by means of remote sensing, as mentioned above (Astakhov et al., 1999). Several exposures in this type of shoreline are documented in Mangerud et al. (2001b), where 18 OSL dates are also reported. Although this is the most important type of shoreline for dating and reconstructing Lake Komi (Mangerud et al., 2001a, b), it is not described further here.

Ice marginal deltas constitute the other type of deposits and are important for correlating Lake Komi with the damming ice margin. Our correlations between Lake Komi and named ice marginal deposits have changed over the last few years and therefore will be briefly reviewed before we describe the deltas.

Astakhov et al. (1999), using satellite images and air photos, mapped a 1200-km-long belt of fresh glacial landscapes and end moraines more or less parallel to the Barents Sea coast. The belt consists of individual morainic segments named from west-to-east the Varsh, Indiga, Markhida, Harbei, Halmer and Sopkay moraines. These were originally considered to be roughly contemporaneous and their collective southern limit was for convenience called the Markhida Line (Mangerud et al., 1999). It was concluded that the ice sheet that formed these moraines dammed Lake Komi. However, a number of new OSL dates from sediments below the Markhida Moraine at its type site yielded ages in the range 80–55 ka (Henriksen et al., 2001). This led Mangerud et al. (2001b) to hypothesize that there had been at least two ice advances, the oldest dammed Lake Komi 90–80 ka and the youngest formed the Markhida Moraine at its type site 60–50 ka. To account for the two

Fig. 6. Aerial photo showing Lake Komi shorelines (arrows). The picture shows a long and narrow peninsula with a distinct shoreline on the left (western) side where there was a long fetch. On the right side there was an embayment and the shoreline is less distinct and preserved only in the lower half of the picture. For location see Astakhov et al. (1999).

different groups of OSL ages we assume that the correlation of the Markhida Moraine with the Indiga–Varsh–Pyoza moraines in the west is valid (Lavrov, 1977; Houmark-Nielsen et al., 2001), but that the Markhida Moraine turns northward into the Barents Sea just east of the Pechora Valley (Fig. 5) (Svendsen et al., 2004). If this is correct, the eastern Harbei and Halmer moraines should not be correlated with the 60–50 ka Markhida Moraine, but represent the ice margin that dammed Lake Komi at 90–80 ka.

3.1.1. Ice marginal deltas

Some distinct marginal ridges of the Harbei moraines are found at the site Ileimusyur. Air photo studies indicate that the ridges were glaciotectonically up-thrusted. Distal to the ridges there is a gently sloping

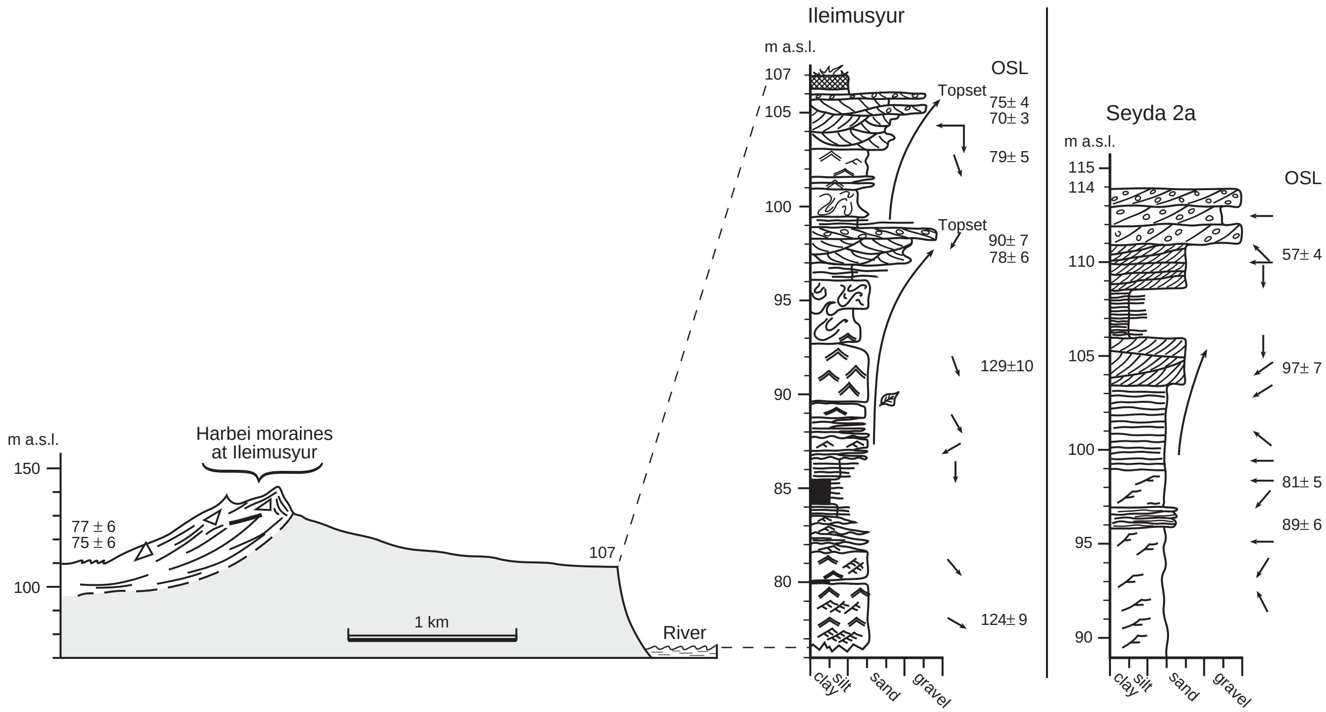

Fig. 7. A schematic cross-section of the Harbei moraines at the site Ileimusyur, including the gently sloping surface interpreted as an ice-marginal delta, drawn from a topographic map 1:200,000 (location in Fig. 5). A log from a section in the distal part of the delta is shown, including obtained OSL dates. Two OSL dates from fluvial sediments proximal to the moraine is plotted from Henriksen et al. (2003). To the very right is a log from the section Seyda 2, for location see Fig. 5.

surface, gradually levelling out and ending in a 30-m-high river bluff, where sand interpreted as distal delta sediments is exposed (Fig. 7). We assume the delta was deposited by meltwater from the same glacier that formed the end moraines, mainly because of their proximity to the site. In the upper part of the exposure there are two gravel units interpreted as wave-influenced top sets. The OSL samples from the shallow water facies yielded typical Lake Komi ages (Table 1, Fig. 7). The two dates from the deeper beds yielded higher ages, but considering that the samples were collected from laminated sand with climbing ripples, deposited at a considerable water depth, we think they were not fully zeroed.

The moraines at Ileimusyur can be followed eastward as a boundary between different landforms, but there the ice limit is not sharply defined by morainic ridges. A shallow meltwater channel is traceable from the ice margin due south to the Seyda 2 section (Fig. 5). Another very distinct meltwater channel ends in the Seyda River valley upstream of that site. At Seyda 2 there is a 25-m high exposure of a deltaic sequence dominated by fine sand (Fig. 7). Three OSL dates yielded ages 97–81 ka, whereas the date from the sand and gravel bed capping the sequence yielded 57 ka. This bed was in the field interpreted as delta topsets. However, as the OSL date indicates it is younger, and because younger fluvial sediments are found at this level elsewhere in the area, we consider that it possibly is younger than the delta. We conclude that the Ileimusyur and Seyda 2 sequences were deposited as glacifluvial deltas in Lake Komi by meltwater streams from the ice sheet that also formed the Harbei moraines. This conclusion is supported by dates from deposits located proximal to the moraines (Fig. 7) (Henriksen et al., 2003).

It is interesting to note that both deltas extend below the present river level, indicating that the valley floors were lower than at present at the onset of delta deposition. Ileimusyur is located in the central part of the more than 150-km-long Rogovaya morainic horseshoe, and probably the deep valley was a result of erosion by that ice advance. We have earlier suggested that the Rogovaya (and Laya-Adzva) moraines represent a slightly earlier phase of the same main advance that formed the Harbei moraines (Mangerud et al., 2001b), an assumption which is supported by the described relationships. According to our photogeologic mapping, Seyda 2 site is located just inside the Rogovaya moraines, invoking a similar explanation.

The deltas demonstrate that a considerable amount of meltwater came out of the southern ice margin.

3.1.2. The outlet of Lake Komi

Possible outlets from Lake Komi were discussed by Maslenikova and Mangerud (2001). The passes through

Table 1
OSL dates

Risø lab no.	Field No.	Depth (m)	Site	Sediment, comments	Age (ka)	Paleodose (Gy)	(*n*)	Dose rate (Gy/ka)	w.c. (%)
972505	96–2020	3	Garevo	Sand bed in the beach gravel.	99±13	104±12	7	1.06±0.05	27
972506	96–2023	8	Garevo	Laminated fine sand.	82±10	142±15	3	1.74±0.09	27
972507	96–2024	7.5	Garevo	Ripple-laminated fine sand.	94±8	168±11	4	1.78±0.09	27
982505	96–2093	1.7	Byzovaya Ravine	Massive sand bed in gravel.	77±8	41±3	11	0.53±0.03	27
982506	97–3083	5.7	Byzovaya Ravine	Cross-stratified fine sand.	54±4	87±4	23	1.61±0.08	27
992526	96–2092	7	Byzovaya Ravine	20 cm thick bed of cross-stratified sand.	110±11	98±8	15	0.89±0.04	27
982507	97–1005	11	Byzovaya Ravine	15–20 cm thick sand. Grades laterally to gravel.	85±7	62±3	32	0.72±0.04	27
982508	97–1003	12.3	Byzovaya Ravine	Parallel laminated fine sand.	90±9	91±7	56	1.01±0.05	27
982517	97–34	1.4	Bolotny Mys	Cross-stratified medium sand.	93±22	104±23	14	1.11±0.06	27
992527	97–35	2	Bolotny Mys	Ripple laminated fine to medium sand	168±13	169±8	15	1.01±0.05	27
982518	97–36	3	Bolotny Mys	Oscillation ripple-laminated fine sand.	68±6	90±6	15	1.34±0.07	27
982516	97–29	5.9	Bolotny Mys	Cross-stratified medium sand.	83±13	104±14	13	1.24±0.06	27
982515	97–28	7	Bolotny Mys	Oscillation ripple-laminated fine sand.	84±14	123±20	36	1.46±0.07	27
982519	97–40	2.6	Ozornoye	Horizontally bedded medium sand.	95±10	112±10	15	1.18±0.06	27
982520	97–45	3.8	Ozornoye	Fine sand. Horizontal lamination and oscillatory ripples.	97±18	140±24	9	1.44±0.07	27
982521	97–52	5.8	Ozornoye	Fine sand with oscillatory ripples.	99±8	140±7	15	1.41±0.07	27
982522	97–59	5	Novik Bosch	Sand below the beach gravel in the N-end of the pit.	98±8	192±9	17	1.97±0.10	27
982523	97–61	9	Novik Bosch	Sand below the beach gravel in the S-end of the pit.	109±8	206±9	12	1.89±0.09	27
992505	98–3075	3	Seyda 2a	Sand in crossbedded sand and gravel.	57±4	67.8±1.8	21	1.20±0.06	27
002519	99–1201	12	Seyda 2a	Crossbedded sand.	97±7	118±5	21	1.22±0.07	27
992504	98–3072	15	Seyda 2a	Medium sand. Laminated.	81±5	112±2	15	1.39±0.07	27
992503	98–3066	18	Seyda 2a	Medium sand. Laminated.	89±6	120±3	15	1.35±0.07	27
002536	99–4217	1.5	Ileimusyur	Sand bed in upper gravel.	75±4	87±2	26	1.17±0.04	20
002537	99–4218	2	Ileimusyur	As sample 4217, collected 0.5 m deeper.	70±4	90±2	23	1.29±0.05	26
002538	99–4219	4.5	Ileimusyur	Fine sand, ripples.	79±5	151±5	21	1.90±0.07	27
002532	99–4210	9	Ileimusyur	Crossbedded sand in lower gravel unit.	90±7	79±4	24	0.88±0.05	23
002533	99–4211	9.1	Ileimusyur	As sample 4210. Collected 10 cm deeper.	78±6	71±3	27	0.92±0.05	22
002534	99–4213	16	Ileimusyur	Fine sand, climbing ripples.	129±10	225±14	21	1.75±0.06	27
002535	99–4215	29	Ileimusyur	Fine sand, climbing ripples.	124±8	214±8	21	1.73±0.07	27
012583	01–196	35	Sangompan, Ob	Well sorted, lacustrine sand	93±7	72±2	21	0.78±0.05	27
012584	01–197	35	Sangompan, Ob	Well sorted, lacustrine sand	82±5	131±4	24	1.61±0.07	31
012544	00–418	7	Yerkata 1, Yamal	Aeolian sand. 19 m a.s.l.	59±4	106±5	36	1.78±0.07	29
012543	00–416	11	Yerkata 1, Yamal	Aeolian sand. 15 m a.s.l.	65±5	118±7	30	1.81±0.07	29
012542	00–414	15	Yerkata 1, Yamal	Lower lacustrine sand. 11 m a.s.l.	72±5	115±6	30	1.59±0.07	33
012541	00–413	17	Yerkata 1, Yamal	Lower lacustrine sand. 9 m a.s.l.	63±4	114 ±5	30	1.83±0.07	34

The first 29 samples are from Lake Komi sediments in the Pechora lowland. The last six samples are from West Siberia (Fig. 10). N gives number of estimates of the paleodose. W.c is water content, given as weight water relative to weight dry matter. Note that age estimates of some earlier published dates are changed due to improved estimates of the contribution of cosmic rays on the dose rate.

the Urals and across the Pechora-Volga watershed are too high. According to satellite imagery and air photo mapping Lake Komi instead extended through the low Tsilma Pass in the Timan Ridge and connected with a lake at the same level in the White Sea Basin (Figs. 2 and 5) (Astakhov et al., 1999; Mangerud et al., 2001a; Nikolskaya et al., 2002). If this interpretation is correct there was one common outlet for Lake Komi and its continuation in the White Sea basin. Buried channels from the White Sea Basin across the water divide to the Baltic Sea are interpreted to represent this outlet (Maslenikova and Mangerud, 2001). However, as this area was affected by the Late Weichselian advance of the Scandinavian Ice Sheet, the channels are buried by till and thus known only from coring and subsurface mapping (Kalberg et al., 1970; Kapishnikova et al., 1994).

3.1.3. Age of Lake Komi

The OSL dates of the Lake Komi sediments are crucial for the Weichselian history of entire northern Russia, because the hypothesis of a major ice-advance and large ice-dammed lakes about 90–80 ka is based almost entirely on these dates (Mangerud et al., 2001a, b).

We have obtained 29 OSL dates from Lake Komi sediments (Fig. 7). The stratigraphic position and other details for the 18 dates from the wave-dominated shorelines are documented in Mangerud et al. (2001b). However, improved estimates of the contribution of cosmic rays on the dose rate have changed the age estimates on some of these samples by as much as 7 kyr because the locally produced dose rate is exceptionally low. We therefore report the corrected ages for previously published samples in Table 1. From the two ice marginal deltas described above we have obtained 11 new dates, also listed in Table 1.

If we discard the outlier of 168 ka from Bolotny Mys and the date of 57 ka from the Seyda 2 site, the unweighted mean for the remaining 27 samples is 92 ± 13 ka, whereas the weighted mean is 82 ± 1.3 ka. The reason for the younger weighted mean is the larger standard deviation for the older samples than for the younger ones. Excluding also the two oldest dates from Ileimusyur, which we consider reasonable, would give unweighted and weighted means of 90 ± 11 and 80 ± 1.3 ka, respectively.

The OSL dates from the Lake Komi sediments show a considerable spread in age, which could, in principle, reflect a long sedimentation history. However, the ages do not increase downward in the stratigraphic succession (Table 1) (Mangerud et al., 2001a). In fact the distribution of the dates (Fig. 8, inset) is very close to the Gaussian distribution predicted by the standard deviations of the individual dates, if all samples were of the same age. The absence of any significant overdispersion in the distribution strongly suggests that the lifetime of the lake was short compared to the uncertainties in the dates, and so it cannot have been longer than a few thousand years.

A general concern for luminescence dating of sediments is the possibility that during deposition the particles were not exposed to sufficient daylight to completely remove any residual dose from previous erosion/deposition cycles. This would lead to incomplete bleaching or zeroing, and thus to a date that is too old. However, this is much less of a problem for the OSL signals, used in this study, than for TL. From a review of OSL ages with independent age control, Murray and Olley (2002) have concluded that incomplete bleaching is unlikely to give rise to significant systematic errors in wind or waterlain sediments older than about 20 ka, except in unusual circumstances (their review included glacio-proximal sediments). This conclusion can to some degree be tested using the Lake Komi sediments, because there are so many samples from different sites seemingly dating the same event. An OSL age is obtained by dividing the paleodose by the dose rate. For the dated samples the measured present day dose rates range from 0.53 to 1.97 Gy/kyr, i.e. they vary by a factor of 4 (Table 1). There is a well-defined linear relationship between present dose rates and the paleodoses, indicating an age of 100–80 ka (Fig. 8). Had incomplete zeroing been a significant factor at some sites, this relationship would show pronounced scatter towards higher paleodoses, with the best bleached samples defining a linear lower limit to the data set. Such a distribution is not seen, suggesting that incomplete bleaching is not a significant factor in these dates, consistent with the conclusions reached by Murray and Olley (2002). Uncertainties in average

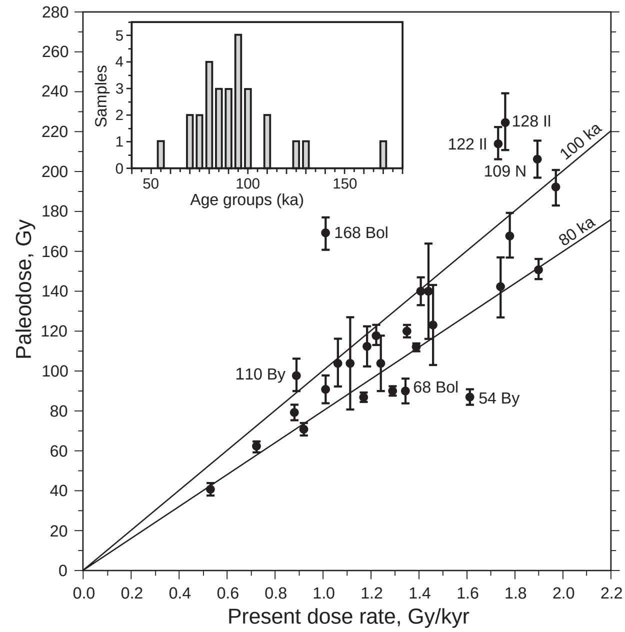

Fig. 8. A plot of paleodose versus present day dose rate for Lake Komi samples. Note that although the dose rate varies with a factor of nearly 4, there is a linear relation between paleodose and dose rate, indicating that the paleodose is a result of the age of the sample and a dose rate close to the present dose rate. Age lines for 80 and 100 ka are indicated. All samples are listed in Table 1. Some outliers are marked with age and site name: Bol = Bolotny Mys, By = Byzovaya Ravine, Il = Ileimusyur, N = Novik Bosch. Inset: A histogram of all OSL dates of Lake Komi sediments (Table 1) showing number of samples in age classes. Only the face value of each date is used, ignoring standard deviations. Except for some outliers, most dates are clustered 80–100 ka.

water content could also give rise to systematic errors in the age. The dose rates used in the age calculations (and Fig. 8) are calculated using the assumption that the sediments were water saturated throughout burial. If the water content was lower for long periods, the dose rate would have been higher; a 1% decrease in average water content would lead to ~1% decrease in age. Saturated water contents are typically ~27%, and we consider it very unlikely that water contents could ever drop much below 10% for well buried sediments. Given that it is the time averaged water content that is relevant, some ages may thus have been overestimated by a few thousand years.

Present dating technology does not allow us to improve the age constraints beyond the uncertainties given, but the mean values given above indicate that the lake most probably existed between 90 and 80 ka, possibly as young as 70 ka. On the other hand most dates we have previously reported from Eemian marine beach sediments in the Pechora Lowland underestimated the known age by ~10% (Mangerud et al., 1999), suggesting a possible systematic underestimation of age by OSL dating. However, these Eemian sediments were dated some time ago with older, less well-developed techniques; we are now in the process of dating a number of recently collected Eemian beach samples using the techniques used in this paper, as a test of the accuracy of the present method.

3.2. The White Sea Basin

The reconstruction of a large ice-dammed lake in the White Sea Basin at 90–80 ka is controversial. The "QUEEN-team" that worked along the major rivers east of the White Sea maintains that no large lake existed at that time (Houmark-Nielsen et al., 2001; Lyså et al., 2001; Kjær et al., 2002). They describe glacilacustrine sediments below till along Pyoza, an eastern tributary to Mezen River (Fig. 1), OSL-dated to 110–90 ka, which they consider could have been deposited in a lake dammed by the ice sheet which dammed Lake Komi (Houmark-Nielsen et al., 2001). However, they conclude the lake was restricted to the upper part of the Pyoza River valley.

On the other hand, as mentioned above, the "QUEEN-team" that worked in the Pechora Basin used air photoes to map Lake Komi shorelines through the Tsilma Pass in the Timan Ridge (Figs. 2 and 5) (Astakhov et al., 1999; Nikolskaya et al., 2002). The implication is that a lake existed at the same level in the White Sea Basin. On the west side of the Timan Ridge the shorelines and sediments were probably destroyed by an ice advance that reached the Syurzi moraines, as discussed below (Fig. 5).

Although the Scandinavian Ice Sheet intruded into the eastern White Sea during the Late Weichselian and eroded most of the underlying pre-Late Weichselian sediments, there are a number of sites where Eemian and Early to Middle Weichselian sediments are preserved on the Kola Peninsula. This is particularly the case along the coast of the White Sea where Eemian marine mud with in situ shells, traditionally related to the Boreal Transgression, serves as a stratigraphical marker horizon (Ikonen and Ekman, 2001, and references therein).

Sixteen sections in the southeastern coastal areas of the Kola, between the rivers Varzuga and Ponoy (Fig. 1), were studied by a third "QUEEN-team". In most of the sections there are glacilacustrine silt, clay and fine sand rhythmites with abundant IRD material and waterlain diamicton found above the Eemian marine beds. These units of deep water facies are unconformably overlain by deltaic and fluvial sands which, in turn, are capped by a Late Weichselian till (Lunkka et al., 2001b). The glacilacustrine sediments are in places more than 15 m thick and their tops are positioned between 5 and 111 m a.s.l. (Fig. 9). In order to determine the minimum age of the glacilacustrine sediments, sand units immediately overlying them were OSL dated at eight sites. Most dates indicate that the fine-grained glacilacustrine units at lower elevations (between 0 and 42 m a.s.l.) were deposited in the White Sea Basin before ca 90–80 ka (Fig. 9). This basin might then have been a part of a huge ice lake connected with Lake Komi.

The above observations and arguments lead us to conclude that Lake Komi probably extended into the White Sea Basin at a level of about 100 m a.s.l. Such a high lake can only be formed if an ice sheet blocked the mouth of the White Sea. As no ice sheet margin is mapped for that time, an arbitrary damming zone across the mouth is used in Fig. 2.

3.3. The West Siberian Plain

To the east of the Pechora Basin, the ice margin that blocked Lake Komi is traced by means of end moraines around the northern tip of the Ural Mountains and across the West Siberian Plain (Astakhov et al., 1999; Mangerud et al., 2001b; Svendsen et al., 2004) (Fig. 2). This ice front must also have blocked the northward flowing drainage of Siberia, including the rivers Yenissei and Ob, thereby causing the formation of a huge ice-dammed lake which overflowed to the Aral and Caspian seas ca 90–80 ka (Mangerud et al., 2001a). A comprehensive review of this drainage history is presented elsewhere (Astakhov, submitted).

Thick glacilacustrine rhythmites lying below 60 m a.s.l. and not covered by tills have long been known in the Yenissei and Ob valleys close to the Arctic Circle (Arkhipov and Lavrushin, 1957; Lazukov, 1970). These deposits predate sediments with non-finite radiocarbon ages and lie stratigraphically above formations of the last interglacial (Astakhov, 1992). The glacilacustrine rhythmites and the above-described moraines were used as arguments to reconstruct a lake up to 60 m a.s.l. at 90–80 ka (Fig. 2) (Mangerud et al., 2001a). Our latest results from the Lower Ob valley support this hypothesis with OSL dates ca 90–80 ka for glacial lake sediments distal to the Sopkay moraines (Fig. 10A).

3.3.1. The outlet of the West Siberian glacial lake and implications for the Caspian Sea

The lowest water divide between the West Siberian Plain and the Aral Sea is the floor of the Turgay Valley, presently at 126 m a.s.l. (Fig. 2). However, the Turgay Valley is filled with thick lacustrine, colluvial and aeolian sediments, radiocarbon-dated to 29–11 ka (Astakhov, submitted), which would be fast removed by river erosion. The bedrock floor of the valley is only 40 m a.s.l. at the present water divide. This elevation was used by Mangerud et al. (2001a), who from that postulated that the West Siberian lake at ca 60 m a.s.l. was dammed by the threshold between the Aral and Caspian seas. However, they had missed that the bedrock floor rises northward from the Turgay Valley to 55 m a.s.l. some 500 km north of the present water divide (Fig. 2) (Astakhov, submitted). This bedrock high

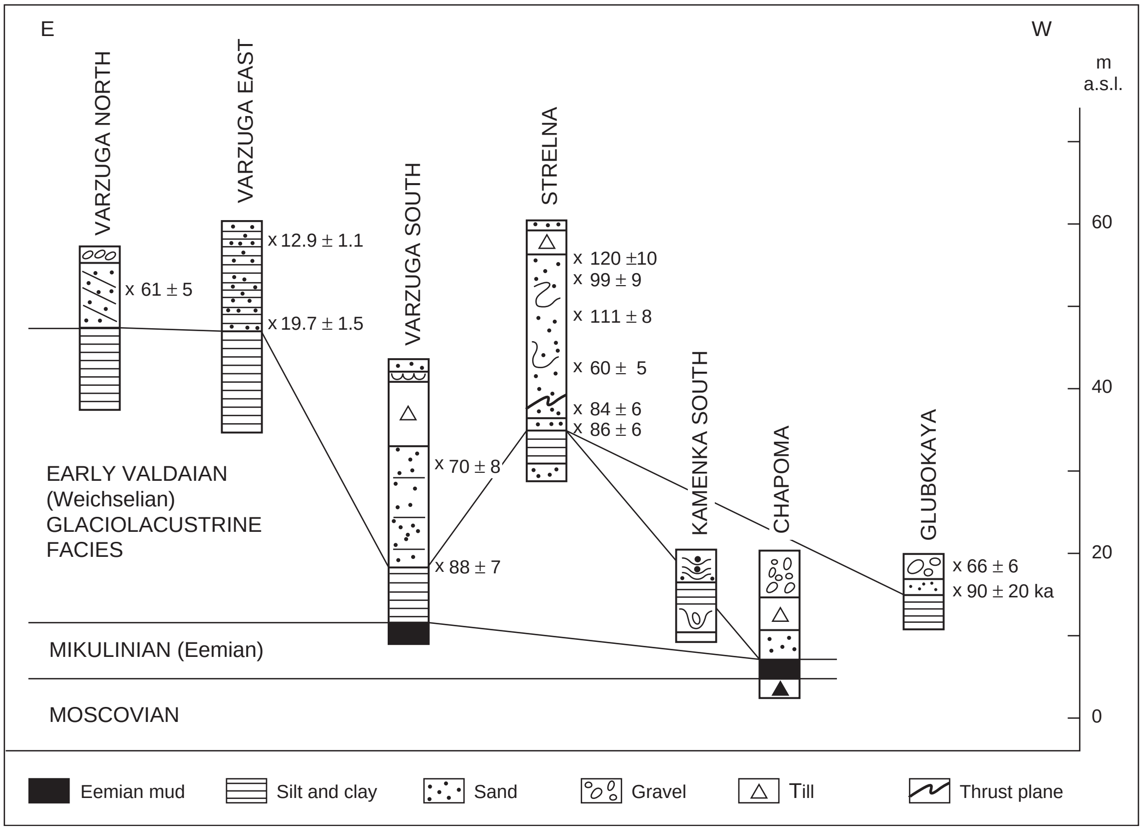

Fig. 9. Sections along the south-east coast of the Kola Peninsula. OSL dates are plotted from Lunkka et al. (in prep).

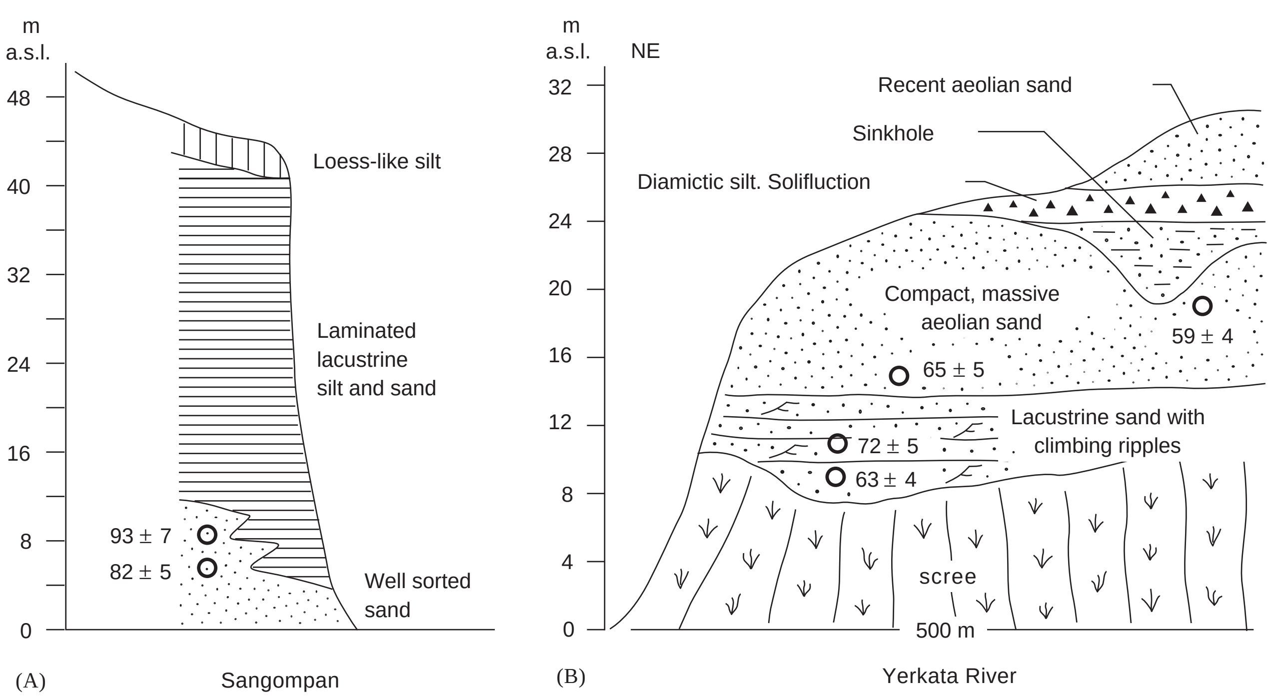

Fig 10. Sections in West Siberia (marked in Fig. 1), with OSL dates plotted. To the left the site Sangompan in the Ob Valley, where lacustrine sediments are referred to the Early Weichselian ice-dammed lake. To the right a section along the Yerkata River on Yamal, where the lacustrine sediments are referred to the Middle Weichselian lake.

probably acted as the dam for the West Siberian glacial lake. From there the water spilled over to the Aral Sea basin and subsequently to the Caspian Sea along the today dry Uzboy channel (Fig. 2).

An overflow from the ice-dammed lake must have influenced the level of the Caspian Sea, as discussed by Mangerud et al. (2001a). Its present level is −28 m, and the altitude of the Manych Pass to the Black Sea is +26 m a.s.l. This latter level does, however, relate to the overflow during the Late Weichselian (Leonov et al., 2002), whereas elevations and ages of older overflows are still poorly known (Zubakov, 1992; Mangerud et al., 2001a).

3.4. Taimyr

The Late Pleistocene glacial maximum on the Taimyr Peninsula occurred in the Early Weichselian. The Barents–Kara Ice Sheet then penetrated 200–250 km south of the Byrranga Mountains, to reach the central parts of the North Siberian Lowland (Kind and Leonov, 1982; Hjort et al., 2004; Svendsen et al., 2004).

When the ice front had retreated to the north of the Byrranga Mountains a glacial lake, dammed by the ice-front, filled the valleys of the Lower Taimyra River and its tributaries. In the beginning the sea might have briefly inundated these valleys, entering through the Byrrangas from the south. The ice front reached a new grounding-line at the North Taimyr ice-marginal zone (NTZ) (Kind and Leonov, 1982; Isayeva, 1984; Alexanderson et al., 2001, 2002). The shoreline was then at 140–120 m a.s.l., as documented by wave cut terraces and deltas along the end moraines. This lake has been OSL-dated by two samples from an ice-contact fluvial deposit, to ca 80 ka (Alexanderson et al., 2001; Svendsen et al., 2004). Glaciolacustrine sediments, often varved, are widely distributed distally to the deltas, indicating a fresh-water regime during the main part of this event (Alexanderson et al., 2001, 2002).

The north-directed normal drainage was reversed and the water from the glacial lake drained southward through the Byrranga Mountains into the Lake Taimyr basin (Alexanderson et al., 2001). Its onward route from there is not yet known, but probably it flowed across the threshold between the Lake Taimyr basin and the Laptev Sea, presently between 60 and 70 m a.s.l. The height difference between the mapped shoreline and this threshold is probably at least partly an effect of glacioisostatic recovery.

Glacilacustrine sediments aged 80–70 ka are found along the present Kara Sea coast north of the Taimyra River mouth (Funder et al., 1999). This is ca 100 km up-ice from the NTZ and these deposits probably derive from glacial damming during the further retreat of the ice front.

4. Middle Weichselian Lakes and rerouting of rivers

4.1. The White Sea Basin

The southernmost Weichselian ice limit in the White Sea Basin is probably the Syurzi moraines (Fig. 5) (Lavrov, 1975, 1977; Lavrov et al., 1991; Svendsen et al., 2004). The configuration of these moraines indicates they were deposited by ice flow upslope from NW, i.e. by the Barents Ice Sheet (Chebotareva, 1977; Lavrov, 1977; Nikolskaya et al., 2002), and not by a local Timan ice cap as suggested by Demidov et al. (2004). This ice sheet dammed lakes as high as 145 m a.s.l. in the upper Mezen Valley (Lavrov, 1975; Demidov et al., 2004). When the ice margin retreated from the Syurzi moraines toward north, there was probably a large ice-dammed lake in front of it. According to the stratigraphy given by Houmark-Nielsen et al. (2001) and Kjær et al. (2002) this must have happened in the early Middle Weichselian. Therefore, the strandlines observable on air photos west of the Timan Ridge at approximately the same elevation as Lake Komi (Astakhov, 1999) should be younger than Lake Komi, in contrast to the correlation of Mangerud et al. (2001a). Their age probably corresponds to the age of glacilacustrine sediments along the Pyoza River, OSL-dated to 70–60 ka if some outliers are excluded (Houmark-Nielsen et al., 2001).

The Varsh–Pyoza moraines (Fig. 5), correlated with the Markhida Moraine in the Pechora Lowland (Lavrov, 1977; Astakhov et al., 1999; Houmark-Nielsen et al., 2001; Nikolskaya et al., 2002), are located to the north of, and are younger than, the Syurzi moraines. Farther west the Pyoza moraines are obliterated by the Late Weichselian advance of the Scandinavian Ice Sheet, but probably the Barents–Kara Ice Sheet at that time blocked the White Sea entrance (Demidov et al., 2004).

Along the southern coast of the Kola Peninsula two sites are known where delta foresets and distal foreset sands of Middle Weichselian age are preserved (Lunkka et al., 2001b). At the Ust Pyalka site climbing ripple and cross-bedded cosets of sand, overlying glacilacustrine silt at around 110 m a.s.l., were OSL dated to 55 ka. In addition, in the Varzuga area delta foreset beds at 50–60 m a.s.l. yielded an age of 60 ka, whereas more distal foreset beds were dated to 68 ka (Fig. 9). There are also other sand units that have yielded Middle Weichselian ages, but their stratigraphical position remains unclear due to deformation by Late Weichselian ice advances.

The ice-sheet reconstructions and OSL dates of glacilacustrine sediments imply that a glacial lake existed in the White Sea Basin at least once during the Middle Weichselian. In the Tsilma Pass (Fig. 5) there is a narrow channel cut into the wider strait formed by the

Lake Komi shorelines. This channel was formed by an overflow from west towards east (Nikolskaya et al., 2002) and is the most probable candidate for the outlet of the Middle Weichselian lake.

4.2. The Pechora Lowland

The ice sheet that formed the Markhida Moraine across the Pechora River Valley probably dammed a narrow lake that filled the valley south of the moraine. The only candidate so far for the sediment from this lake is the black clay on the Sula River at 30–55 a.s.l. which overlie terrestrial sand with OSL ages of 74–53 ka obtained by older techniques (Mangerud et al., 1999). A flat sandy depression that could have acted as an outflow channel is identified east of the Pechora Valley (Nikolskaya et al., 2002). The highest threshold of this depression is 60–70 m a.s.l., and an overflow would have been directed to the eastern Pechora Sea and farther to the Kara Sea. A lake of this elevation would be restricted to the lower part of the Pechora Valley.

4.3. The Kara Sea and the West Siberian Plain

If the reconstructed ice-sheet configuration is correct (Fig. 3), it would imply an ice-dammed lake in the southern part of the Kara Sea, penetrating into the low-lying parts of the plain. The only potential outlet leads eastward across the North Siberian Lowland to the Laptev Sea (Fig. 3), but the location of an overflow channel is unknown to us. According to the GLOBE digital terrain model the lowest water divide is between 40 and 50 m a.s.l. As a hypothesis to test in the future, we have constructed a lake at 45 m a.s.l. (Fig. 3). Sediments that we believe were deposited in this lake are found in a section along the Yerkata River on southern Yamal. Here lacustrine sand is OSL dated to 72–63 ka, an age supported by OSL dates at 65–59 ka from the overlying aeolian sand (Fig. 10B).

4.4. Taimyr

During the Middle Weichselian the Barents–Kara Ice Sheet again reached the NTZ and cut off the normal northward drainage of the Taimyra River. Glacial lakes were dammed in the lower reaches of that river and its tributaries. As during the preceding stage, the water drained southward into the Lake Taimyr basin. Ice-front deltas and more distal shorelines document a lake level not higher than 80 m a.s.l. It is dated to ca 60 ka by 7 OSL dates (Alexanderson et al., 2001, 2002; Svendsen et al., 2004). In the map (Fig. 3) this lake is connected with the above-discussed reservoir of West Siberia at 45 m a.s.l.

5. Late Weichselian Lakes and rerouting of rivers

5.1. Taimyr

No ice-dammed lakes have been documented along the Last Glacial Maximum (LGM) ice margin on the Taimyr Peninsula. However, such lakes might have existed in the lowest parts of the river valleys. The reconstructed ice sheet (Fig. 4) must have reversed the drainage pattern, again leading to an outflow southward through the Taimyra River Valley into the Lake Taimyr basin. This is probably reflected by an increase in both water level and rate of lacustrine sedimentation in the Lake Taimyr basin around 19 ka (Möller et al., 1999; Niessen et al., 1999). The further course of the spillway is not documented, but the 25 m a.s.l. threshold to the Pyasina River basin is a likely candidate for an outlet to the west.

5.2. Did the Barents–Kara Ice Sheet dam any major lake?

The extent of the Barents–Kara Ice Sheet is not confidently mapped in the eastern part of the Kara Sea. We find it most probable that the extent of the LGM ice was limited and therefore allowed rivers to freely discharge into the Arctic Ocean along the eastern margin (Stein et al., 2002; Svendsen et al., 2004). An alternative hypothesis visualises a narrow ice dam across the northern Kara Sea that extended onto the Taimyr Peninsula (Fig. 4) (Alexanderson et al., 2001, 2002; Polyak et al., 2002). The latter would imply at least a temporary blockade of the drainage with an ice-dammed lake formed in the Kara Sea south of the dam. Sediments supporting this hypothesis are possibly found on the floor of the Kara Sea (Polyak et al., 2002).

No lacustrine sediments of LGM age are reported from sites above present sea level along the shores of the Kara or Pechora seas. Instead, thick coarse silts with syngenetic ice wedges of LGM age are described down to sea level on the Yamal Peninsula (Vasil'chuk et al., 1984, 2000). These "Yedoma-type" sediments are most likely ice-rich aeolian silt (Astakhov, 1992). Furthermore in central West Siberia only loess-like silts changing laterally into thaw-lake sediments are recognised in Late Weichselian sediments below 40 m a.s.l. (Astakhov, 1992). On the Barents Sea coast of the Pechora Lowland aeolian sand, OSL-dated to about 21 ka, occur down to present-day sea level (Mangerud et al., 1999, 2002). Bottom silts of the Yenissei Estuary yielded a wide range of AMS dates from 20 ka to the present (Kuptsov and Lisitsyn, 2001).

5.3. The eastern flank of the Scandinavian Ice Sheet

The palaeohydrology during the Late Weichselian was dramatically different from that of the Early and

Middle Weichselian in northern Russia-Siberia. The Barents–Kara Ice Sheet did not at this time block the northbound Russian rivers (Fig. 4). However, the Scandinavian Ice Sheet attained its greatest extent during the Late Weichselian at ca 18–17 ka (OSL dates and thus calendar yrs), and glacial lakes extended some 500 km along the ice margin (Fig. 4). These lakes, of which the highest shorelines today are at 130 m a.s.l., drained to Volga and eventually into the Caspian Sea (Kvasov, 1979; Lunkka et al., 2001a). A main transgression and overflow of the Caspian Sea is dated to about 19–17 ka (16–14 ka ^{14}C-years) (Leonov et al., 2002).

When the Scandinavian Ice Sheet started to retreat from its Late Weichselian maximum position at ca 17–15 ka, a new series of short-lived glacial lakes was formed along its margin. The palaeohydrology was convincingly worked out by Kvasov (1979), but timing of the events was uncertain. Based on the position of the ice margin during the Vepsian and Luga stages, it is suggested that the meltwater first drained through the southern Peribaltic area toward central Poland and via "Urstromtäler" to the Elbe River. At about 16–15 ka (this and younger ages partly based on varve chronology) the ice margin had retreated so far that the water drained into the southern part of the Baltic Basin. When the ice retreated from the Luga moraines at ca 14.2 ka an initial Baltic Ice Lake was formed in the southern part of the basin, shown in Fig. 4. An ice-dammed lake known as the Privalday Lake, which occupied extensive areas south of Lake Ladoga and Gulf of Finland, drained to this Baltic Ice Lake across southern Estonia and central Latvia (Fig. 4).

An ice lake in the Onega basin existed between 14.4 and 12.9 ka (Saarnisto and Saarinen, 2001). It was mainly fed by meltwater from the east where the Vodla River drained extensive ice-dammed lakes. The outlet of the Onega glacial lake was across the Onega–Ladoga isthmus into the Baltic (Fig. 4). This lake phase terminated when the ice lake emptied to the White Sea basin, following the retreat of the ice margin in the north. The current outlet of Lake Onega, River Svir, towards Lake Ladoga and the Baltic, originated more than 10.5 ka (Saarnisto et al., 1995). During the entire Late Weichselian the Onega basin remained above sea level and thus no sea connection existed between the White Sea and the Baltic basins (Saarnisto et al., 1995), contrary to interpretations of some earlier workers (Sauramo, 1958).

5.4. Areas and volumes of the lakes

The areas and volumes of the main paleolakes discussed above are given in Table 2. The numbers are calculated for the outermost ice-front position according to the maps (Figs. 2 and 3). During ice-margin withdrawal, the areas and volumes of the lakes probably increased considerably as the lake followed the retreating ice margin. Therefore, estimating an impact of the water outburst, the volumes in Table 2 should rather be taken as minimal.

5.5. Modelling the outburst of the ice-dammed lakes

We have modelled the outburst of Lake Komi because we consider it the best mapped and dated of the ice-dammed lakes discussed in this paper. However, we consider that the results can be qualitatively applied also for the other lakes. Estimates of the magnitude of the final outburst flood from Lake Komi have been obtained by applying the physically based outburst flood model described by Clarke (2003). The model assumes that the flood is released through a subglacial tunnel in a similar manner to modern outbursts such as those from the Grímsvötn reservoir in Iceland (Björnsson, 2002) and is a reworking of the Spring and Hutter (1981) generalisation of Nye's (1976) theory. The model simulates the spatio-temporal evolution of water pressure, water velocity, water temperature and cross-sectional area of a semi-circular drainage conduit fed by an upstream reservoir. Many of the model inputs correspond to well-known physical properties of ice and water (Clarke, 2003). The main sources of uncertainty in the model are lake geometry, subglacial flood routing, lake level relative to sea level at the time

Table 2
Some properties of the paleolakes

Ice dammed lakes	Lake level (m a.s.l.)	Area ($\times$ 1000 km^2)	Volume ($\times$ 1000 km^3)	Average depth (m)
Lake Komi in the Pechora Lowland, 90–80 ka	110–90	76	2.4	32
White Sea Basin, 90–80 ka	100	218	15	69
West Siberian Plain, 90–80 ka	60	610	15	24
Taimyr, 80 ka	140	45	2	45
Total 90–80 ka			34.2	
West Siberian Plain, 60–50 ka	45	881	32	36
Taimyr, 60 ka	80	17	0.4	25

Lake level is given relative to present day sea level. The numbers for the 90–80 ka lakes are from Mangerud et al. (2001a). The numbers for the West Siberian lake at 60–50 is calculated here, and the numbers for the Taimyr lakes are from Alexanderson et al. (2001). For the White Sea Basin we postulate a lake at 60–50 ka of the same magnitude as at 90–80 ka.

of the flood, water temperature in the lake and hydraulic roughness of the drainage channel.

The hypsometric geometry of the reservoir has been calculated using the modern topography (GLOBE-Task-Team, and-others, 1999), with the lake level given by observed shorelines of Lake Komi. For outburst flood modelling the relevant hypsometric function is for the available water rather than the total water, a distinction that depends on the flood routing and size and location of basins of trapped water that develop as the lake drains. The lake level was taken as 150 m a.s.l. level at that time (Fig. 2). The flood path was assumed to follow the present valley of the Pechora River, then extended into the Barents Sea (Fig. 2). This is the most probable route along the topographic lows, but the length of the path is poorly constrained. For sensitivity studies we have selected three paths (A, B, C in Fig. 11) that have lengths L of 373, 638 and 834 km, respectively. The shortest path is considered the most likely. For all models we assume a uniformly spaced grid of 51 points.

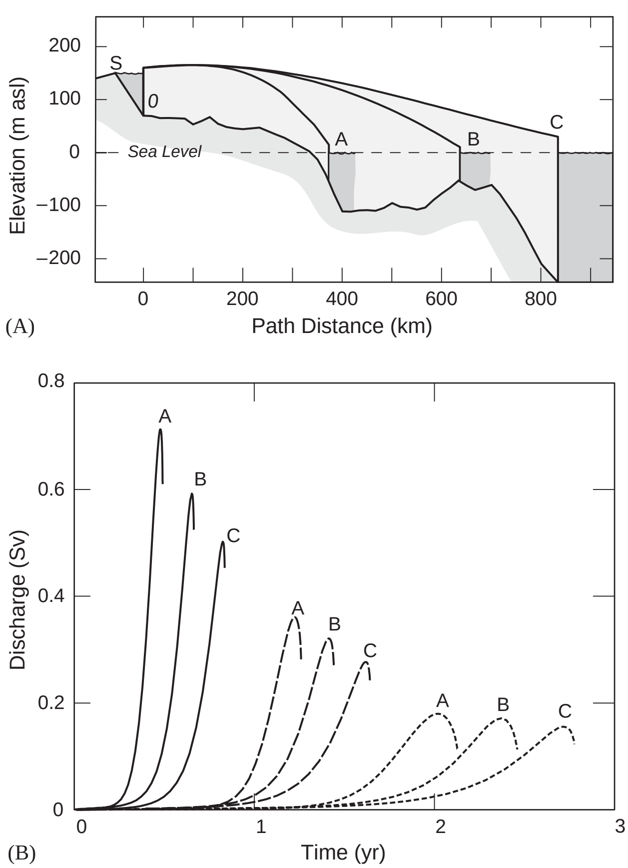

Fig. 11. (a) Geometric details for the three outbursts flood paths considered. The alternative drainage routings are labelled as A (373 km), B (638 km) and C (834 km). Bed topography is based on modern topography. The lake surface is 150 m above contemporaneous sea level. Profile A is the preferred flood geometry and yields the most intense floods. (b) Simulated flood hydrographs for a suite of model runs. Routes are labelled A, B and C as in (a). Lake temperature is indicated by solid curves (4°C), long-dashed curves (2°C) and short-dashed curves (1°C). The preferred temperature is 4°C.

Although unknown, the ice topography is tightly constrained by the bed topography and by the necessary condition for outburst flood release—that the ice overburden pressure must be close to the subglacial water pressure when the flood is initiated. For this reason the ice surface elevation cannot greatly exceed the initial lake level and therefore the maximum ice elevation along the flood path is taken as 165 m a.s.l. at 90 ka. Because the model results are not sensitive to fine details of ice or bed topography we have simply assumed a cubic form for the ice surface topography $Z_s(s) = Z_0 + As + Bs^2 + Cs^3$. The constants Z_0, A, B and C are chosen so that the surface elevation was near the flotation elevation at the endpoints $s = 0$ and L, with an ice divide at an elevation of 15 m above the pre-flood lake level (hence 165 m a.s.l. at 90 ka) and 100 km downstream from the channel inlets (Fig. 11a).

Other important controls on flood magnitude are the temperature of the lake and hydraulic roughness of the channel. Lake water temperature can only be inferred. Because the maximum density of water is at 4°C this is a common temperature for lakes in cold environments and is our preferred temperature for Lake Komi. For comparison we also consider water temperatures of 1°C and 2°C but consider especially 1°C to be improbably cold. An empirical measure of the hydraulic roughness of natural and artificial drainage channels is provided by the Manning roughness parameter. This is not readily measurable but can be estimated for modern outburst floods. Clarke (2003) performed this analysis and found that for three well-studied floods the implied Manning roughness lies in the range 0.023–0.32 $m^{-1/3}$ s, which is consistent with estimates for rivers and streams (e.g. Dingman, 1984, p. 144). For the simulation modelling we take $n' = 0.30\,m^{-1/3}$ s as the reference value, but bracket this from below and above by also considering $n' = 0.25$ and $0.35\,m^{-1/3}$ s.

Results of the flood simulations are summarised in Fig. 11b. Not surprisingly the shortest routing with the highest lake temperature yields the most intense flood of 0.71 Sv, whereas the longest routing with the lowest temperature yields the mildest flood of 0.16 Sv. For the maximum flood, peak discharge increases from 0.71 to 0.85 Sv if the roughness is decreased to $n' = 0.25\,m^{-1/3}$ s and the discharge decreases to 0.31 Sv if the roughness is increased to $n' = 0.35\,m^{-1/3}$ s. For all floods the lake volume is 2100 km^3, although the integrated volume of the flood hydrographs ranges from 2170 km^3 for the maximum flood to 2450 km^3 for the minimum flood. Because we assume an inflow to the lake of 4500 $m^3\,s^{-1}$, floods of long duration receive more of this inflow volume. As the preferred model values are the 373 km drainage path and 4°C lake temperature, our best estimate of the maximum discharge from the 90 ka outburst from Lake Komi is in the range 0.3–0.85 Sv with a duration of 2–3 months, depending on which value of Manning roughness is selected. Without better knowledge of their geometry and filling level, it would be premature to perform comparable simulations for the other ice-dammed lakes discussed in this paper. Assuming that the lake temperature and hydraulic roughness are the same for all the floods, the major influences on flood magnitude are the reservoir volume and the average slope of the flood routing. The lakes in the White Sea Basin and West Siberian Plain have estimated volumes that are six-fold that of Lake Komi (Table 2) and it seems probable that the flood magnitudes would correspondingly exceed that for Lake Komi and the flood durations would be comparable.

6. Discussion

6.1. Uncertainties in the reconstruction of the lakes

The concept of blocking the northbound Russian rivers by ice sheets is simple and has been known for decades. Yet, as pointed out above, our factual knowledge from geological records is often limited. Reasons for this include the facts that the lakes extended over enormously large and remote areas, and that until recently dating of sediments older than the limit of the ^{14}C method was almost impossible.

The history of the lakes on Taimyr is reasonably well established. However, the only well-mapped ice-dammed reservoir in the large lowlands is Lake Komi in the Pechora Lowland, OSL-dated to 90–80 ka. We have postulated a contemporaneous and much larger lake on the West Siberian Plain as an implication of the ice sheet reconstruction. This hypothesis has still to be tested, because the glaciolacustrine sediments in that area are yet not comprehensively dated (Fig. 10A). The postulated contemporaneous lake in the White Sea basin is better supported by OSL dates, but its actual existence is disputed (Houmark-Nielsen et al., 2001) and further field studies and dating efforts are necessary to decide which alternative is correct. Nevertheless, we believe that the largest ice-dammed lakes existed during the Early Weichselian, about 90–80 ka (Table 2 and Fig. 2).

Unfortunately, except for Taimyr, geological records of the Middle Weichselian proglacial system are even less known than for the Early Weichselian. However, if the ice-sheet reconstruction for 60–50 ka is approximately correct, there should have been a major lake in the White Sea basin. A few OSL dated and some undated lake sediments support this interpretation. There were probably also a small lake in the Pechora Valley and a large lake over parts of the present sea-floor of the Kara Sea and the lower parts of the West Siberian Plain. However, such lakes are not documented by shorelines, and so far only a few dates are available from West Siberia (Fig. 10).

For the Late Weichselian we are confident that no sizeable lakes existed above present sea level along the coasts of the Kara and Pechora seas, and only a small and short-lived proglacial lake, if any, could have formed on the Kara Sea shelf.

The main picture is clear for the Scandinavian Ice Sheet. The east- and northbound drainage was only blocked during the Late Weichselian glacial maximum (LGM). At that time proglacial water was forced across the continental watershed into the Volga River and farther to the Caspian Sea.

6.2. Drainage to the Caspian and Black Seas

We have identified only two periods when drainage was diverted toward the Caspian and Black Seas. Interestingly, it was two different drainage areas that were affected alternately.

The first time the drainage of the West Siberian Plain, including the Yenissei and Ob rivers, was forced towards the Aral, Caspian and Black Seas by the advancing Barents–Kara Ice Sheet (Fig. 2). We consider the damming and rerouting certain, but the dating is imprecise. However, we correlate this event with the existence of Lake Komi in the Pechora Lowland, OSL-dated to 90–80 ka, an age supported by two dates from West Siberia (Fig. 10A).

The second time the drainage was re-routed southward was during the LGM, about 18–17 ka. This time the Scandinavian Ice Sheet blocked the drainage toward the Baltic Sea and meltwater from a considerable sector of the ice sheet flowed over the watershed to the Volga River. This possibly caused the Caspian Sea to overflow along the Manych Pass at 26 m a.s.l. to the Black Sea. According to Leonov et al. (2002) the last overflow from the Caspian Sea lasted from 19 to 13 ka (16–11 ka ^{14}C-years) and was interrupted by a regression at 17–16 ka.

6.3. Impact of lake outbursts to the Arctic Ocean

We have described above our modelling of the outburst from Lake Komi, which indicates that it occurred over a few months. We assume the other lakes also drained over a comparable period of time although the water would follow different pathways. The outbursts of lakes in the White Sea Basin and in the Pechora Lowland would end in the Barents Sea. Whether the fresh water from there would flow toward the Arctic Ocean or directly west into the Norwegian Sea would depend on the glacier configuration and pattern of ocean circulation at that time. However, the ice sheet on the Barents Sea was certainly almost gone when the outburst occurred. The large lake on the West

Siberian Plain and the lakes on Taimyr would be drained to the Kara Sea and further into the Arctic Ocean Basin. In summary, we expect that most of the water would end up in the Arctic Ocean, and follow the Transarctic Drift through the Fram Strait to the Norwegian Sea every time an ice-dammed lake was drained.

Considering that water volumes of these lakes are equal to that of 20–50 years of the present-day runoff, and a much longer period of the dry glacial climate, we expect that several of the outbursts had significant impact on sea-ice formation and other environmental parameters (Mangerud et al., 2001a). Before speculating on these possible impacts we would like to point out that if diagnostic parameters for the outburst of the individual lakes could be found in marine cores, it would provide important stratigraphic markers for land-sea correlations. Candidate signals from outbursts from the discussed ice-dammed lakes are now found in the Arctic Ocean (Svendsen et al., 2004).

The combined final outburst volume from lakes Agassiz and Ojibway, Canada, which has been suggested as the trigger for the 8200 BP cooling event (Barber et al., 1999), was considerably larger (Teller et al., 2002) and released 3–5 times more water than the total volume of the 90–80 ka lakes described here (Table 2). On the other hand, the volume released from the Herman Beach phase of Lake Agassiz, a candidate for the Younger Dryas cooling (Broecker et al., 1989), was 9500 km^3, i.e. considerably less than the volume of several of the individual lakes discussed here. If the total volume (34,400 km^3) of the 90–80 ka lakes were distributed across the Arctic Ocean with continental shelves, it would produce a nearly 4 m thick fresh water layer. Restricted to an area approximately corresponding to the Transarctic Drift, the layer would be more than 22 m, and distributed on the Norwegian–Iceland–Greenland seas, it would amount to a 13 m layer.

It is known from modelling studies that modern thermohaline circulation (THC) in the North Atlantic is sensitive to freshwater perturbations (Bryan, 1986). In fact, the freshwater budget in the North Atlantic is considered the main governor of the strength of the Atlantic THC; dynamic models show that it is sensitive

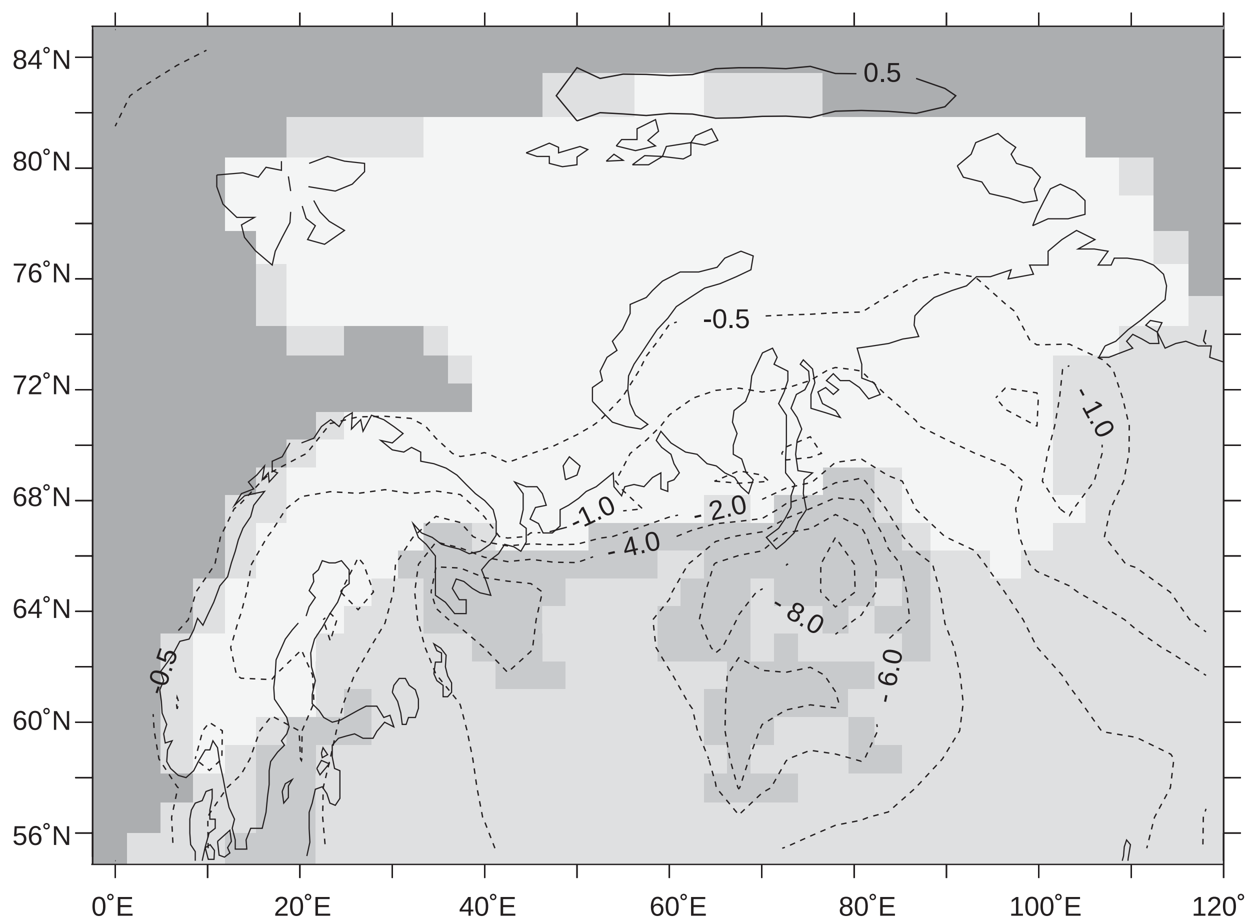

Fig. 12. Impact of the lakes on summer (June, July, August) air temperatures at about 100 m above the surface (in °C). Contour lines are −10°C, −8°C, −6°C, −4°C, −2°C, −1°C, −0.5°C, and +0.5°C (negative contour lines dashed), given as the difference between the model run with lakes and the run without lakes. The underlying shading indicates the surface type (from dark grey to white: ocean, lakes, continental soil, ice sheet). Thick line: present-day coastline.

to even small changes on the order of 0.1 Sv (Stocker and Wright, 1991). The THC could completely collapse due to an increased freshwater input (Schmittner and Clement, 2002) and it has been suggested that this took place during large glacier melt-water events. From the above and from our modelling results indicating that Lake Komi alone would provide in the range of 0.3–0.85 Sv, it is tenable to suggest that outbursts of the ice dammed lakes in northern Russia could have triggered cold events during Early and Middle Weichselian deglaciations that may have been comparable to Younger Dryas. One can also speculate that the fresh water pulse at 50–60 ka in the Norwegian Sea (Hebbeln et al., 1998) was a result of the outburst of the 50–60 ka lakes described here.

6.4. Climatic impacts of the lakes on the continent and ice sheet

In addition to the resultant feedback to continents from lake outbursts, which changed ocean circulation, the actual presence of large lakes on continents will influence the regional climate (Bonan, 1995; Lofgren, 1997). Concerning the climatic impact of ice-dammed lakes, Hostetler et al. (2000) showed that Lake Agassiz cooled the climate near the southern limit of the Laurentide Ice Sheet 11,000 years ago, leading to reduced precipitation rates over the ice sheet. Krinner et al. (2004) quantified the climatic impact of the Early Weichselian ice-dammed lakes described in this paper. They used the polar version of the LMDZ3 (Lab. de Météorologie Dynamique, CNRS, Paris) stretched-grid general circulation model (Krinner et al., 1997) which includes a thermal lake module (Krinner, 2004). Two simulations of the Early Weichselian climate, one with and one without ice-dammed Russian lakes, were carried out.

The results show that the ice-dammed lakes, because of the cool climate near the ice sheet and because they are fed by fairly large amounts of meltwater from the southern margin of the ice sheet, remain cool in summer. In particular, the lakes are significantly cooler than the normal continental soil, with which they are replaced in the control simulation. As a consequence, the lakes exert a cooling effect aloft. The cooling is strongest over the lakes themselves, but it extends onto the adjacent continent including the ablation zone of the ice sheet (Fig. 12). Compared to a situation without ice-dammed lakes, summer melt on the southern ice sheet margin is therefore strongly reduced, and the ice sheet mass balance is increased. This means that there was a positive feedback loop involving the ice sheet and the proglacial lakes: ice sheet growth allowed the existence of the lakes, which in turn favoured further growth of the ice sheet. A similar amplifying effect must have existed during the ice sheet retreat. Therefore the lakes play an important role in the dynamics of the Eurasian climate during this period. Other climatic effects of the lakes include a slight modification of the seasonal cycle of precipitation on the ice sheet.

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