Marginal formations of the last Kara and Barents ice sheets in northern European Russia

VALERY I. ASTAKHOV, JOHN INGE SVENDSEN, ALEXEI MATIOUCHKOV, JAN MANGERUD, OLGA MASLENIKOVA AND JAN TVERANGER

BOREAS

Astakhov, V. I., Svendsen, J. I., Matiouchkov, A., Mangerud, J., Maslenikova, O. & Tveranger, J. 1999 (March): Marginal formations of the last Kara and Barents ice sheets in northern European Russia. *Boreas,* Vol. 28, pp. 23–45. Oslo. ISSN 0300-9483.

Glacial landforms in northern Russia, from the Timan Ridge in the west to the east of the Urals, have been mapped by aerial photographs and satellite images supported by field observations. An east–west trending belt of fresh hummock-and-lake glaciokarst landscapes has been traced to the north of 67 °N. The southern boundary of these landscapes is called the Markhida Line, which is interpreted as a nearly synchronous limit of the last ice sheet that affected this region. The hummocky landscapes are subdivided into three types according to the stage of postglacial modification: Markhida, Harbei and Halmer. The Halmer landscape on the Uralian piedmont in the east is the freshest, whereas the westernmost Markhida landscape is more eroded. The west–east gradient in morphology is considered to be a result of the time-transgressive melting of stagnant glacier ice and of the underlying permafrost. The pattern of ice-pushed ridges and other directional features reflects a dominant ice flow direction from the Kara Sea shelf. Traces of ice movement from the central Barents Sea are only discernible in the Pechora River left bank area west of 50°E. In the Polar Urals the horseshoe-shaped end moraines at altitudes of up to 560 m a.s.l. reflect ice movement up-valley from the Kara Ice Sheet, indicating the absence of a contemporaneous ice dome in the mountains. The Markhida moraines, superimposed onto the Eemian strata, represent the maximum ice sheet extent in the western part of the Pechora Basin during the Weichselian. The Markhida Line truncates the huge arcs of the Laya-Adzva and Rogovaya ice-pushed ridges protruding to the south. The latter moraines therefore reflect an older ice advance, probably also of Weichselian age. Still farther south, fluvially dissected morainic plateaus without lakes are of pre-Eemian age, because they plunge northwards under marine Eemian sediments. Shorelines of the large ice-dammed Lake Komi, identified between 90 and 110 m a.s.l. in the areas south of the Markhida Line, are radiocarbon dated to be older than 45 ka. The shorelines, incised into the Laya–Adzva moraines, morphologically interfinger with the Markhida moraines, indicating that the last ice advance onto the Russian mainland reached the Markhida Line during the Middle or Early Weichselian, before 45 ka ago.

Valery I. Astakhov & Olga Maslenikova, Geological Faculty, St. Petersburg University, Universitetskaya 7/9, St. Petersburg 199034, Russian Federation; John Inge Svendsen, Centre for Studies of the Environment and Resources, University of Bergen, Høyteknologisentert (HIB), N-5020 Bergen, Norway; Jan Mangerud & Jan Tveranger, Department of Geology, University of Bergen, Allégt. 41, N-5007 Bergen, Norway; Alexei Matiouchkov, VSEGEI (National Geological Institute), Sredny pr.74, St. Petersburg 199026, Russian Federation. E-mail: valery@NB1466.spb.edu; *received 26th January 1998, accepted 4th January 1999*

The extensively discussed problem of the last glaciation of Arctic Russia (Grosswald 1980, 1993, 1994; Astakhov 1992, 1997, 1998a; Faustova & Velichko 1992; Velichko *et al.* 1997) has several dimensions addressed in three coordinated papers in this volume, presenting the results of six years of work in the northeast of European Russia. The chronological aspect is discussed by Mangerud *et al.* (1999), and a three-dimensional reconstruction of the last ice sheet that affected the Pechora Basin is presented by Tveranger *et al.* (1999). In this paper, we document glacial and periglacial features in the European part of the Russian mainland, which has been mapped in order to improve the material base for reconstructing the Weichselian ice sheet extent east of the area affected by Fennoscandian glaciers. These results are closely connected with the sediments and chronostratigraphy discussed in Mangerud *et al.* (1999), where it is concluded that the Barents and Kara ice sheets probably did not reach the Russian mainland during the Late Weichselian. Consequently, the mapped glacial landscapes described in this paper are of pre-Late Weichselian age. A hypothesis of the offshore location of the Late Weichselian glacial limit is presented by Svendsen *et al.* (1999).

For decades, Quaternary mapping of Northern Russia by local geological surveys has been largely influenced by non-glacial theories of the drift origin, which has made it difficult to use middle-scale maps for the purpose of our research. For a long time the only data available for reconstructing, correlating and modelling former ice sheets were small-scale general maps of Quaternary deposits (Yakovlev 1956; Krasnov 1971).

The first detailed study of glacial landscapes was performed by Moscow geologists led by A. Lavrov, who produced photogeological maps in the scale 1:200 000 for vast areas west of 60°E (Lavrov 1977;

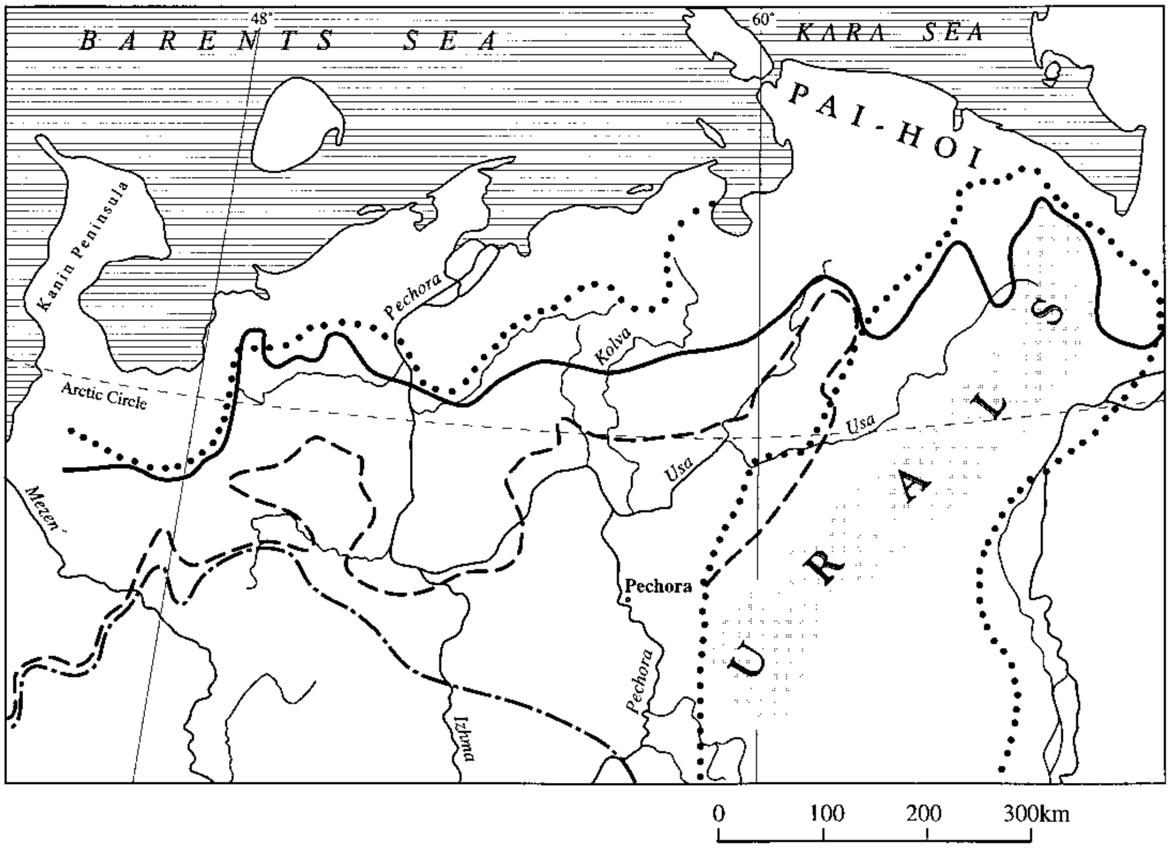

Fig. 1. Map of northern Russia with Weichselian ice-sheet limits according to different authors. Our reconstructed ice-sheet limit corresponds to the southern boundary of Markhida, Harbei and Halmer types of hummocky morainic landscapes (Markhida Line) shown in Fig. 2. (······Yakovlev 1965: Early Weichselian; – – – Lavrov 1977: Late Weichselian; –·–·– Arslanov *et al.* 1987: Late Weichselian; —— Present authors: Early/Middle Weichselian (Markhida Line)).

Lavrov *et al.* 1986, 1991). These data were used for palaeoglaciological models by Grosswald (1980, 1993, 1994), who over the last two decades has advocated the existence of a huge Late Weichselian ice sheet that covered much of northern Russia down to 64°N (Fig. 1). In contrast, another recent hypothesis, using the same photogeological data, depicts a Late Weichselian (or slightly older) ice sheet reaching only 66°N (Biryukov *et al.* 1988; Faustova & Velichko 1992).

Another controversy concerns the location of the major ice domes. In addition to the traditional idea of an ice dome centred over the Urals, Yakovlev (1956) inferred glaciation centres over the Barents Sea and Novaya Zemlya, based on the provenance of erratics and on the fact that the position of the ice sheet margin is parallel to the coast (Fig. 1). This scheme was challenged by Kaletskaya (1962), who maintained that since the bulk of the glacial clasts was represented by Pai-Hoi and Uralian Palaeozoic rocks, only these uplands could have hosted Late Pleistocene ice domes. The concept of ice sheets centred on the Barents Sea shelf reappeared again in the 1970s after the west–east striking marginal belts were mapped in more detail and a gradual decrease of pebble content in the tills to the south was established (Lavrov 1977; Lavrov *et al.* 1986).

Independent of the concept of a shelf-centred glaciation, most authors took it for granted that a major ice dome was situated in the Polar Urals (Yakovlev 1956; Lavrov 1977; Biryukov *et al.* 1988; Faustova & Velichko 1992). The main reason for this assumption was the occurrence of Uralian stones in various tills east and west of this narrow mountain range. However, no geomorphic evidence to support this idea has been presented. Only small loops of alpine moraines have been mapped along the European slope of the Urals, generally not farther than 5–8 km from the mountain front (Gesse *et al.* 1963).

The idea of a major ice dome in the Urals was also at variance with erratics, transported from the Kara Sea coast to the southwest across the Pai-Hoi Range, and with the northeast direction of striae and eskers on the Pai-Hoi and Vaigach Island reported by Voronov (1951) and Tarakanov (1973), who therefore suggested an additional ice dome around the Yamal Peninsula and in the southwestern Kara Sea area. Astakhov (1979) did more detailed observations on the eastern slope of the Polar Urals, supported by photointerpretation around the northern tip of this mountainous range, and found no alpine moraines on the Uralian piedmont north of 68°N. The Palaeozoic bedrock was found polished and striated from the north and covered by west–east striking morainic ridges. These results, incompatible with a Uralian ice dome, served as one of the cornerstones in reconstructions of a large ice sheet centred on the Kara Sea shelf (e.g. Arkhipov *et al.* 1980; Astakhov 1992; Grosswald 1993, 1994).

Methods and principles of mapping

Aerial photographs and satellite images

Our data on surficial glacial features have been derived mostly from stereoscopically studied aerial photographs at basic scales of 1:50 000 and 1:35 000 (Figs. 6–12). We used a total of about 4500 aerial photographs and 60 photomosaics, obtained for the most part by aircraft surveys between 1988 and 1991. These are of better quality than the images of the 1940–1960s interpreted by Lavrov (1977), Lavrov *et al.* (1986, 1991) and Arslanov *et al.* (1987). In addition, high-altitude aerial photographs and Russian high-resolution satellite images of scales 1:150 000 and 1:280 000 were employed for tracing especially lengthy features, or in places where aerial photographs were not available, as in the areas south of the Arctic Circle and west of the Timan Ridge. These satellite images have much better ground resolution (5–8 m) than the Landsat images used by Punkari (1995). For identifying some morainic ridges, such as those shown in Fig. 9 even 1:50 000 aerial photographs are sometimes not detailed enough.

Field observations

Most photogeological objects were compared with and verified by ground observations. The main targets of our field inspections are located along the Pechora, Sula, Shapkina, Kuya, Kolva, Usa and More-Yu rivers, along the Timan coast of the Barents Sea and at some inland localities accessible only by helicopters. Principal test sites are shown in Fig. 2 and partly described in Mangerud *et al.* (1999). The key sections around the city of Naryan-Mar (Markhida, Vastiansky Kon, Kuya River, Fig. 2) have been visited repeatedly over the years to observe, in three dimensions, features exposed differently after each spring flood. Several sites, especially those containing Palaeolithic artefacts, were excavated in a more comprehensive manner. Additional ground checks have been made during short helicopters stops along the Barents Sea coast, to the south of Urdyuga Lake and along the western slope of the Polar Urals. Routes by boat along major rivers and by car along rare paved roads proved to be the most instrumental method for collecting ground evidence and interpolating between the sections studied in detail. Gravel pits and other road excavations, which appeared in the 1970s around the cities of Naryan-Mar, Ust-Tsilma and Usinsk, have given a new insight into the nature of the otherwise poorly exposed periglacial sediments.

Large-scale maps and borehole profiles by local exploration teams from the Arkhangelskgeologia and Polarnouralgeologia corporations have been consulted to evaluate surface lithologies and sediment thicknesses. We have confirmed the interpretation of large photogeological objects as glaciotectonic ridges and ice-contact plateaus given in Lavrov's works.

The Pechora Basin has predominantly weak glacier beds consisting basically of Quaternary and Mesozoic sand and clay. The soft substrate accounts for the rare occurrence of subglacial features such as striae, flutes, tunnel eskers and stoss-and-lee topography. Forms of ice disintegration, such as kames, supraglacial eskers, terraces of confined glacial lakes, occur frequently (cf. Lavrov *et al.* 1991). However, these features are poor indicators of ice-flow patterns and age of ice-sheet advances. Therefore, the main targets of our mapping, aimed at reconstructing the location and succession of ice-sheet margins, have been ice-pushed ridges, marginal and subglacial meltwater features and shorelines of proglacial lakes. Ice disintegration features were mapped collectively, in so far as their assemblages could be used as possible clues to the extent and age of corresponding glaciations.

Differences between glacial and permafrost degradation forms

An important step in our research was to identify sedimentological and geomorphological processes operating in the Late Pleistocene–Holocene in the study area. We have found that the evolution of postglacial landscapes in the Pechora Basin, due to persistent Pleistocene permafrost, has been governed by periglacial processes, as in Siberia, rather than by temperate climate processes, as in postglacial western Europe. According to our field interpretations, many small hummocks and accompanying diamictons are not real signatures of the last ice-sheet advance, as was assumed by, for example, Arslanov *et al.* (1987) and Grosswald (1993), but rather the result of postglacial degradation of the Pleistocene permafrost. A reinvestigation of the key section at Markhida (9 in Fig. 2) has revealed that most of the surficial diamictons are not basal tills, but gravity-driven flow-tills and solifluction sediments produced in the Holocene by repeated topographic inversion of the landscape due to melting of former thick permafrost (Tveranger *et al.* 1995). Similar thick sheets of uncompacted postglacial diamictons, previously interpreted as basal tills of the Upper and Middle Pleistocene, have also been described in many

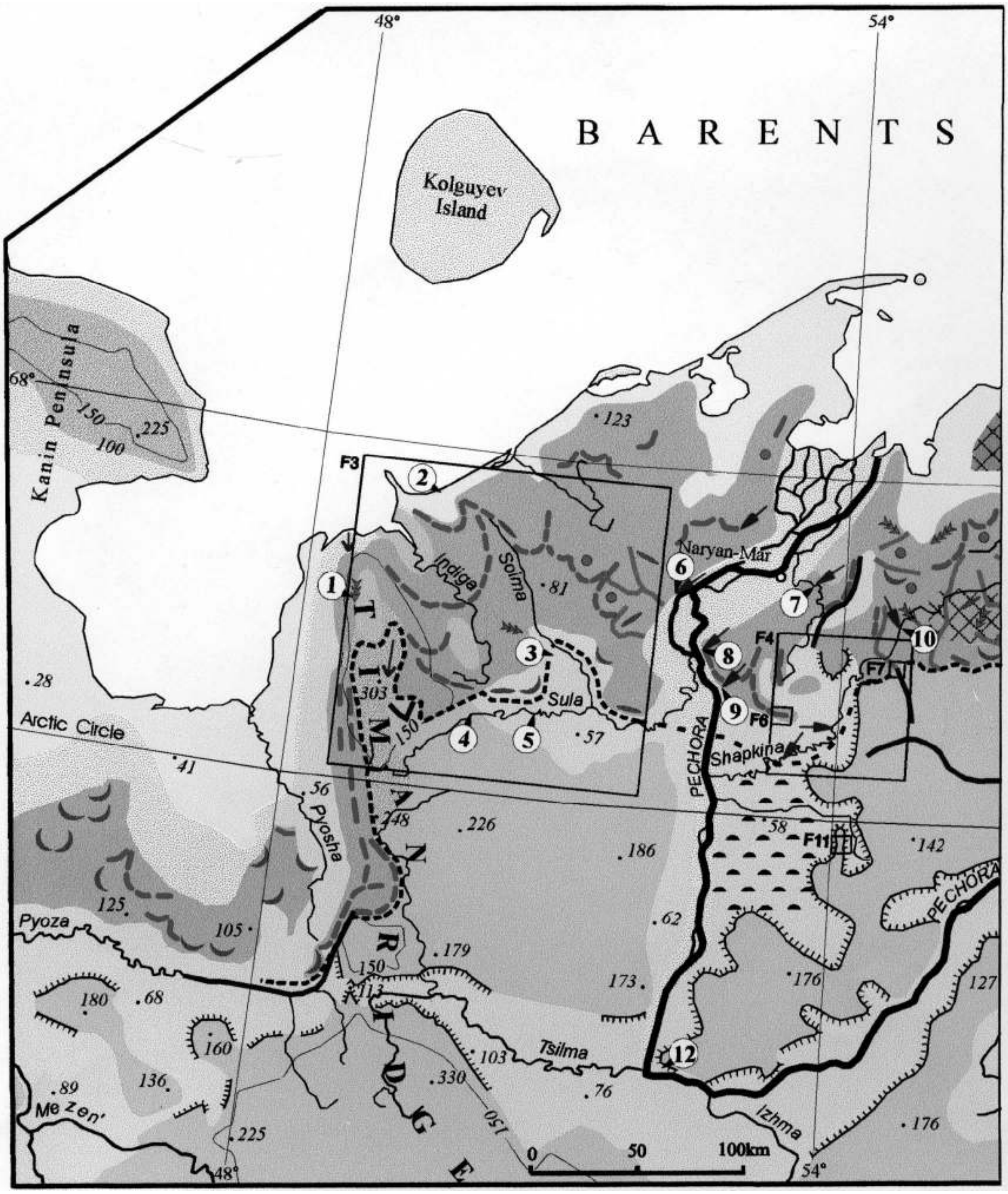

Fig. 2. Map of glacial and periglacial features. Non-studied areas are shown in grey shades, depending on altitudes. Rectangles with numerals relate to figures in the text. The broken line is the inferred ice-sheet limit named the Markhida Line, i.e. the southern boundary of hummock-and-lake landscapes of Markhida, Harbei and Halmer types. Black arrows with circled numerals indicate key sections in Mangerud *et al.* (1999): 1 – Harius Lakes; 2 – Timan Beach; 3 – Urdyuzhskaya Viska; 4 – Sula section 7; 5 – Sula section 22; 6 – Hongurei; 7 – Upper Kuya; 8 – Vastiansky Kon; 9 – Markhida; 10 – Upper Shapkina; 11 – Akis; 12 – Garevo; 13 – Ust-Usa; 14 – Novik; 15 – Ozyornoye; 16 – Bolotny Mys; 17 – Yaran–Musyur; 18 – Haryaha; 19 – Podkova–1; 20 – Yarei–Shor. Palaeolithic sites excavated: 21– Byzovaya; 22 – Pymva-Shor; 23 – Mamontovaya Kurya. Sites on the Yamal Peninsula acording to Gataullin and Forman (1997) and Gataullin *et al.* (1998): 24 – Marresale; 25 – Mutny Mys.

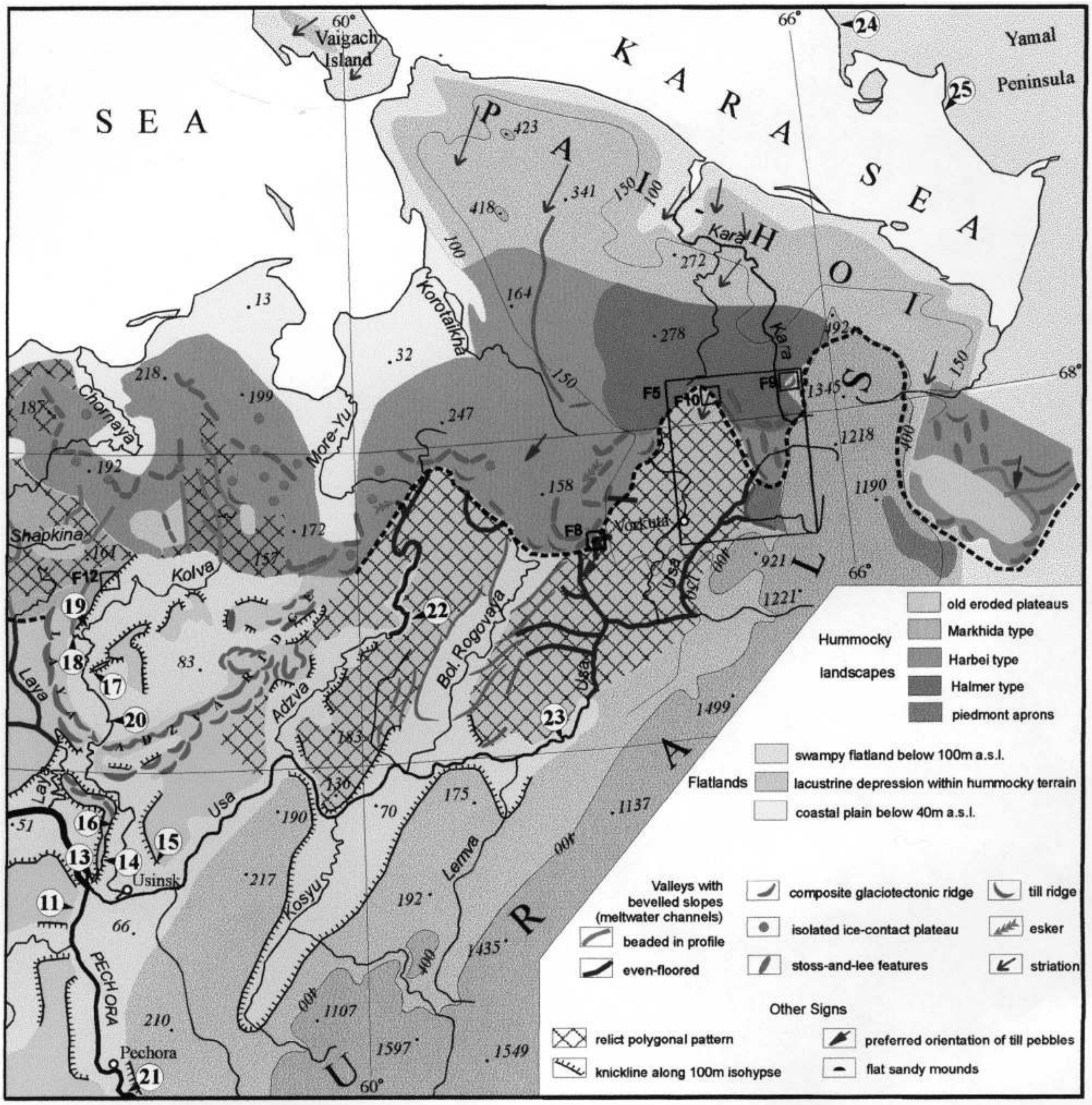

Fig. 2. Continued.

wide clearings along the Sula, Shapkina and Pechora rivers.

Hummock-and-lake landscapes: different types and age

Hummock-and-lake landscapes derived from the melting of stagnant glacier ice and from degradation of the Pleistocene permafrost were previously believed to be morphologically convergent (Boitsov 1961). We have found it possible to qualitatively differentiate between large assemblages of landforms derived from glaciokarst processes and from melting of permafrost. We used such characteristics as shape, size, density and orientation of small lakes and their inverted counterparts, accretion hummocks (Astakhov 1998b). We then considered the ratio of thermokarst landforms versus glaciokarst landforms in a given terrain. This ratio is difficult to quantify, but can be assessed from a range of images typical for different landscapes (Figs. 6–12). In

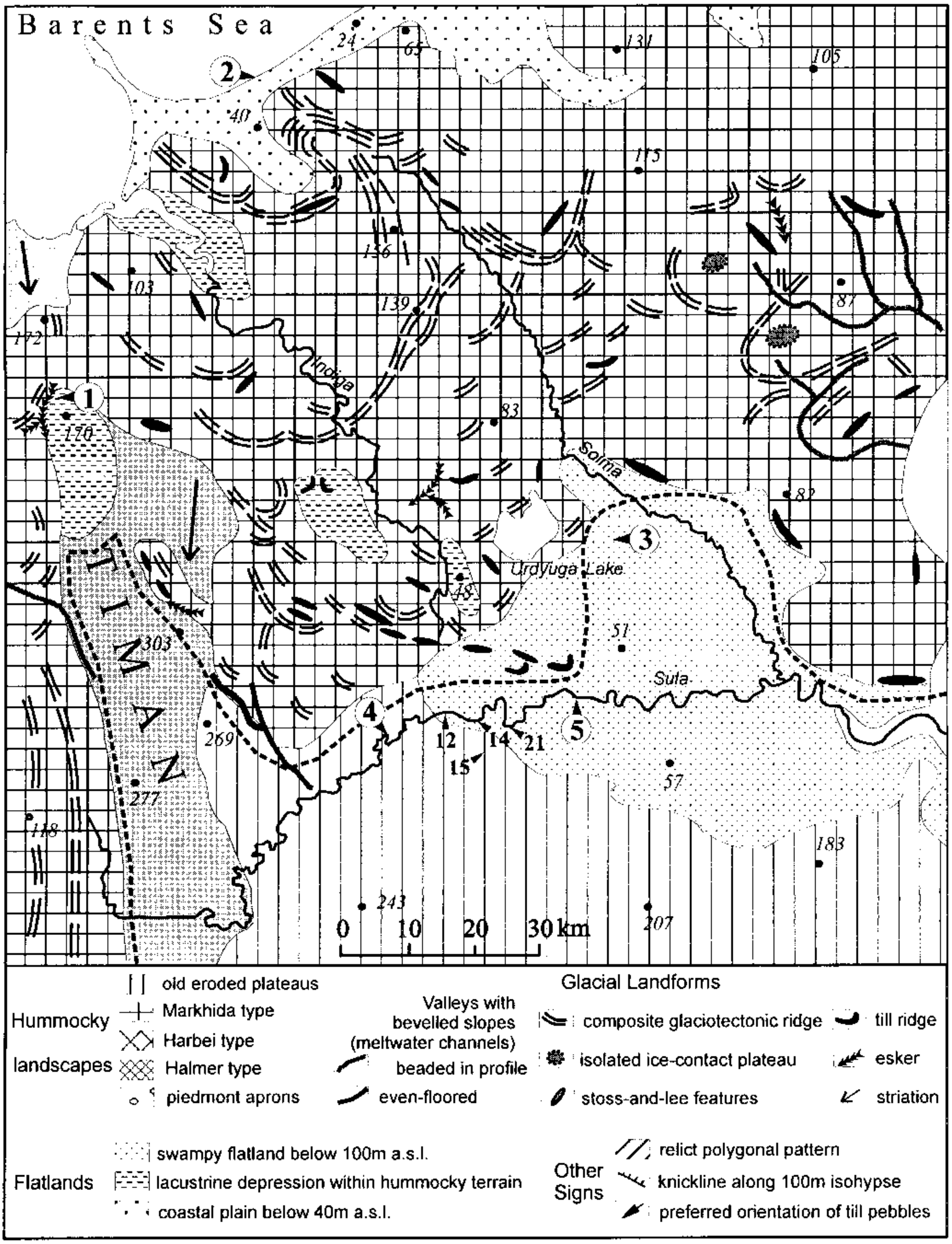

Fig. 3. Photointerpretation map of the Timan–Sula area (see location in Fig. 2). Circled numerals correspond to the site numbers in Fig. 2. Numbered arrows without circles indicate studied exposures. The broken line is the inferred ice-sheet limit corresponding to the Markhida Line.

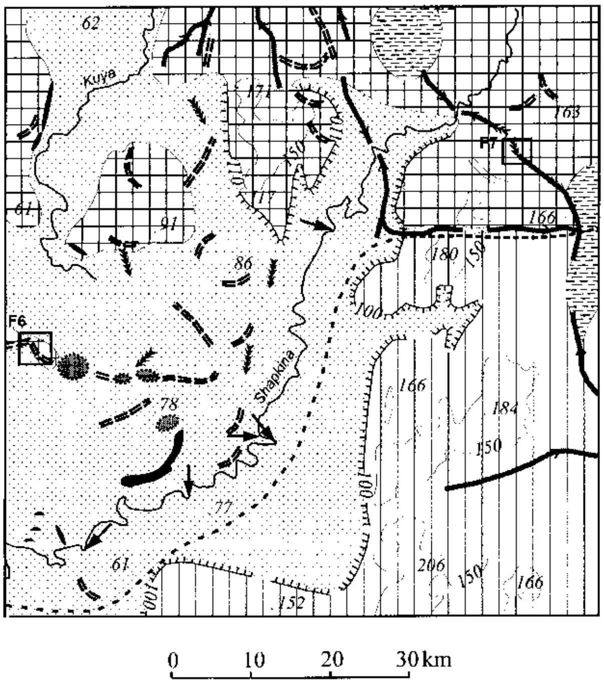

Fig. 4. Photointerpretation map of Shapkina River valley (see location in Fig. 2). Explanation is in Fig. 3. Rectangles are locations of Figs. 6 and 7. The broken line is the inferred ice-sheet limit corresponding to the Markhida Line. Note the mapped shorelines about 100 m a.s.l. that seem to be incised into the Markhida morainic landscape.

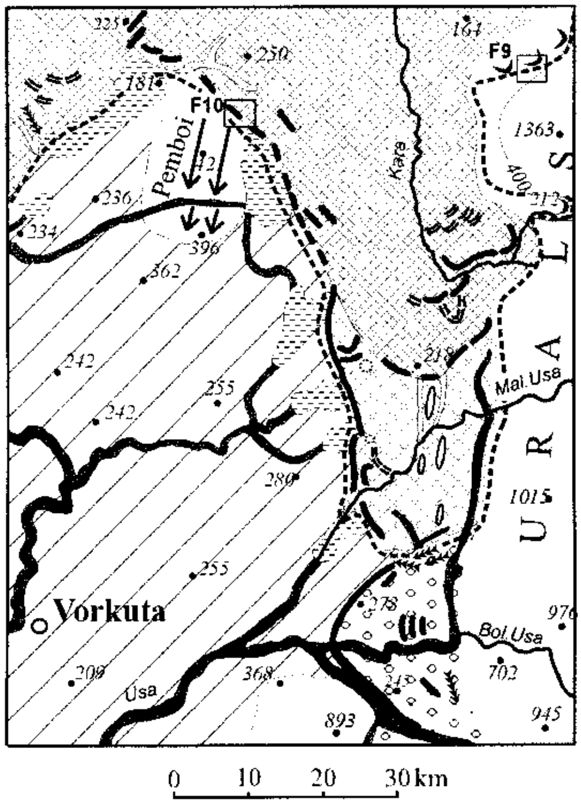

Fig. 5. Photointerpretation map of the western slope of the Polar Urals (see location in Fig. 2). Explanation is in Fig. 3. Rectangles are locations of Figs. 9 and 10. Note that the area inside the reconstructed ice-sheet lobe along the Urals includes two types of landscape, the Halmer type in the north and the Harbei type in the south. The horseshoe-shaped morainic ridge along Bol. Usa River outlines a local piedmont glacier that obstructed southward flow of the ice sheet lobe in the north.

a perennially frozen region this ratio may be used to estimate the morphological age of a glacial landscape, i.e. the degree of its transformation into a landscape dominated by permafrost forms. We also use the density of small lakes as a simple indicator of a stage of glaciokarst/thermokarst development. The number of lakes per square unit reaches a peak soon after the start of a climatic amelioration and then slowly decreases during further degradation of permafrost and/or stagnant glacier ice.

A significant complication in the interpretation of the morphological age of northern landscapes is that they are a function of both the time that elapsed since the ice sheet disintegration and the postglacial climate. The climate, governing the thickness and temperature of local permafrost, would directly affect the number and size of thermokarst lakes and the rate of their evolution, and thereby the morphological diversity of photogeologically mapped landscapes. Therefore, our mapping results should be considered in conjunction with the stratigraphic data described by Mangerud *et al.* (1999).

The glacial landscapes of the region are diverse, ranging from lake-dominated terrains of ice disintegration (Fig. 10) to typical landscapes of fluvial erosion. Our task was to trace the main spatial trends, locate boundaries between different types of landscapes and, where possible, correlate them with changes in sedimentary sequences. In our map (Fig. 2) all variety of glacial topography is reduced to some main landscape types, which may or may not have stratigraphic implications. The boundaries between the landscapes are not distinct everywhere, and at places there are gradational transitions. However, in our opinion the difference between the extreme morphological types is clearly seen, if the southernmost terrains are compared with the northern ones, and if areas along the Pechora River are set against the areas around the Polar Urals (Figs. 7 and 9).

The map (Fig. 2) is a generalized version of our original photointerpretation maps compiled in the scale 1:200 000. Some features are shown in more detail in the maps of specially important areas (Figs. 3–5).

Young hummocky landscapes

In general, the map (Fig. 2) shows an east–west trending belt of relatively fresh-looking glacial landscapes

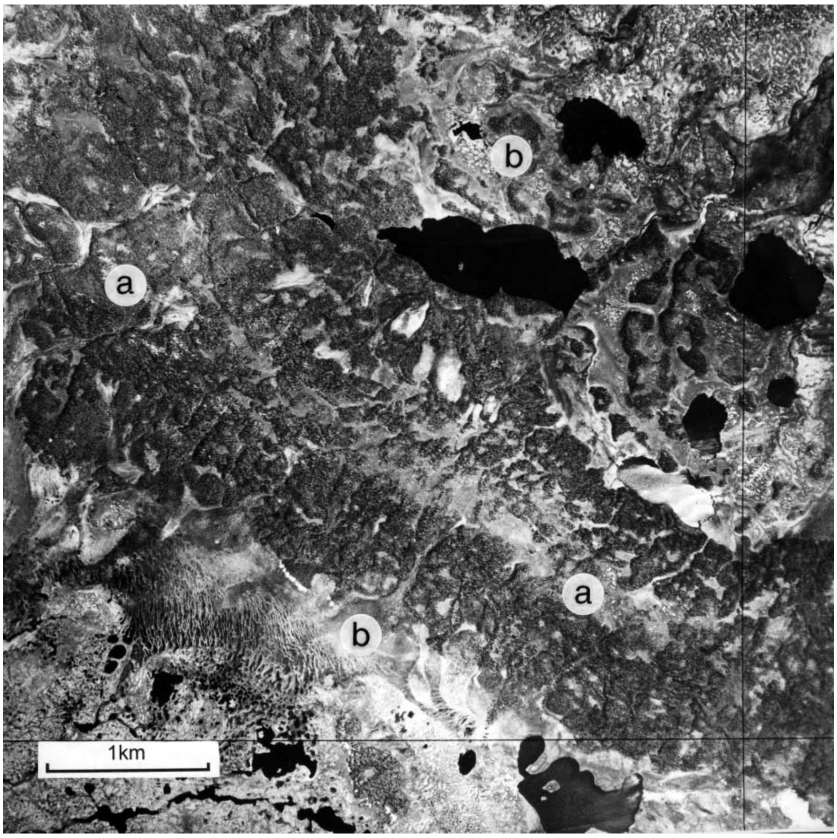

Fig. 6. Aerial photograph of the Markhida marginal ridge (see location in Figs. 2 and 4). (a) – faintly expressed parallel ridges along the forested distal slope. The crest of the ridge at 70–80 m a.s.l. is flat and possibly eroded by a proglacial lake; (b) – swampy flatland at 30–40 m a.s.l., the deepest part of the proglacial Lake Komi.

coloured green. They are characterized by various small accretion hummocks with intervening lakes, morainic ridges, isolated ice-contact plateaus, meltwater channels and rare eskers. This zone comprises several morainic segments, partly described by previous authors. They are the Varsh moraines on the western slope of the Timan Ridge, the Indiga moraines on its eastern slope (Lavrov 1977), the Markhida moraines in the lower Pechora River catchment area (Lavrov 1977; Grosswald 1980) and the Sopkay moraines in the eastern Urals (Astakhov 1979). In our map, we have added the Harbei moraines east of the Markhida moraines and the Halmer moraines west of the Urals. All the moranic landscapes of the Varsh–Indiga–Markhida–Harbei–Halmer–Sopkay belt are readily identifiable in aerial photographs (6–12). Their collective southern margin, called the Markhida Line, is interpreted as the limit of the same ice advance, and is the main morphological boundary in the entire study area. Morphologically, the segments of the morainic belt can be classified in three main types of glacial landscapes: the Markhida type in the west, comprising the Varsh, Indiga and Markhida moraines proper, the Harbei type in the middle, including the Sopkay moraines east of the Urals, and the Halmer type along the western slope of the Polar Urals (Fig. 2). Morainic landscapes of different types are coloured different shades of green: the darkest green corresponds to the freshest (morphologically youngest) glacial landscape.

Hummocky morainic landscape of Markhida type

This landscape has a rugged topography due to steep slopes, kettled depressions, isolated conical hillocks and especially the gentle and wide ice-pushed ridges mapped by Lavrov (1977, 1978). However, the summit surface is normally only 80–100 m a.s.l. Higher morainic plateaus with altitudes over 150 m are less than 10–15 km wide. The fluvial network consists mostly of subparallel consequent valleys draining into major tributaries of the Pechora River or directly into the sea. Most rivers either start at interfluve lakes or connect them.

Surficial permafrost in this landscape is thin (20–100 m) and probably postdates the Middle Holocene. Along the Pechora valley, permafrost is absent altogether. Eastwards, between the upper Shapkina River and the coast, the permafrost layer grows thicker (up to 200 m) and is presumably older (Yershov 1988).

The most conspicuous feature, making this landscape entirely different from the older moraines to the south, is the numerous lakes up to 10 km across and 30 m deep. The area is also dotted with small interfluve ponds (10 000 to 60 000 m^2). We estimate that there are about

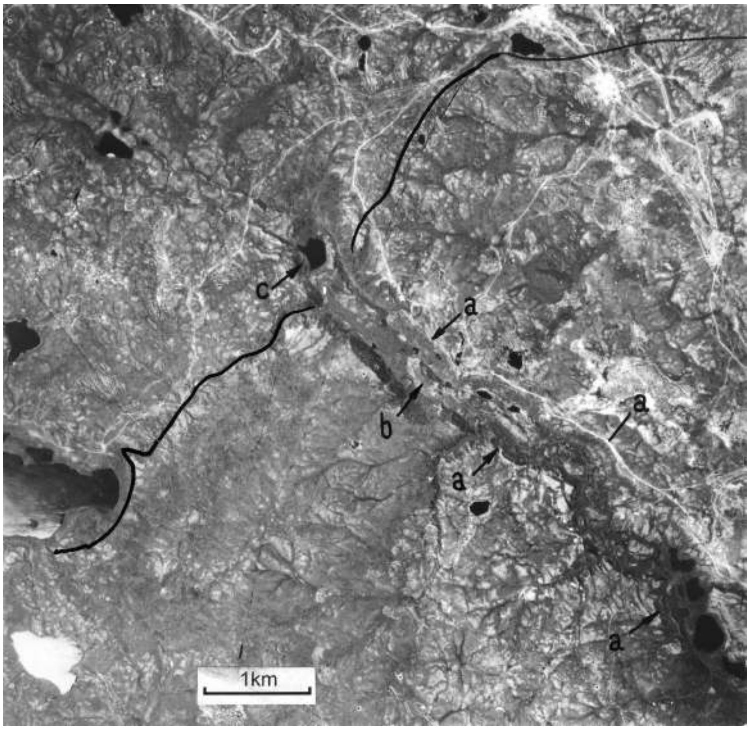

Fig. 7. Aerial photograph of the Markhida Line (solid line) at the left bank of Shapkina River (see location in Figs. 2 and 4). A plateau built of pre-Weichselian tills at 140–160 m a.s.l. east of the Markhida Line is cut by a dry valley, 30–40 m deep, with bevelled slopes. The valley is a continuation of small eskers inside the glacial landscape of Markhida type. (a) – break of the plateau into the meltwater channel; (b) – elongated erosion residuals along the uneven valley floor; (c) – deep funnel-shaped lake, probably a glacial mill between the head of the meltwater channel and the end of an esker to the north.

four ponds per km^2. Whereas large and deep lakes of intricate configuration are interpreted as glaciokarst lakes, the ubiquitous occurrence of many small, shallow (1–3 m deep) and round lakes is attributed largely (but not exclusively) to thermokarst sinking on the present-day permafrost. Many shallow thermokarst lakes are encircled by remnant boggy platforms, locally called hasyrei, showing recent lateral migration of the lake. Small funnel-shaped glaciokarst lakes, which are relatively rare in this landscape (cf. marginal assemblages in Figs. 6, 7 and 12 with Figs. 8 and 10), occur either in confined depressions or atop the higher plateaus.

The higher plateaus composed of thick tills normally bear only isolated patches of hummock-and-lake topography. The gentle slopes of the plateaus are characterized with numerous solifluction tongues with streamlined microrelief, which coalesce into flat aprons at the base (Figs. 6 and 12). Often the solifluction mantle is covered by thick sheets of eolian sand (Mangerud *et al.* 1999).

The most expressive glacial feature are composite ridges, especially west of the Pechora valley, where they make a continuous chain (Figs. 2 and 3). They are crescentic or horseshoe-shaped, 3–8 km wide, 30–80 m high and make up interfluves at 80–130 m a.s.l. The proximal concave slope is usually the steepest. The summit surface has an undulating microrelief due to numerous parallel ridges 1 to 10 m high. These small ridges are typically 0.2–0.5 km long, and their strike line follows the crest of the main ridge, outlining the arc-like configuration. The small ridges are commonly built of steep-dipping sand, whereas the intervening troughs are shaped by solifluction flows along the strike of silt and clay beds. Therefore, the sets of small parallel ridges are interpreted as a result of selective erosion in permafrost environment, the permeable sand being resistant to solifluction. The erosional nature of the parallel ridges is also indicated by deflation armours and conical gravelly residuals along the crests (Astakhov 1979), and by the lack of mantling till on the flattened surface of many large composite ridges. Where the mantling till is preserved, or where the thrusted strata are mostly clay and silt, composite ridges are gentler, being covered by solifluction sheets. This is probably the case east of the Markhida section, where the parallel ridges along the distal slope of the large marginal ridge are barely visible because they are obscured by a solifluction mantle (Fig. 6).

Flat swampy depressions or lakes often occur under the proximal slopes of composite ridges, making hill-hole pairs characteristic of glaciotectonism (Levkov 1980). The glaciotectonic origin of composite ridges is evident in the high bluffs of the Vastiansky Kon section (8 in Fig. 2), where alternating slices of interglacial sand and basal till dip at angles of 30 to 40° to northeast (Tveranger *et al.* 1998a). Structurally, the composite ridges are similar to glaciotectonic imbrications called skibas in the western Russian plain (Levkov 1980), or the arc-like overthrust-injection assemblages of West

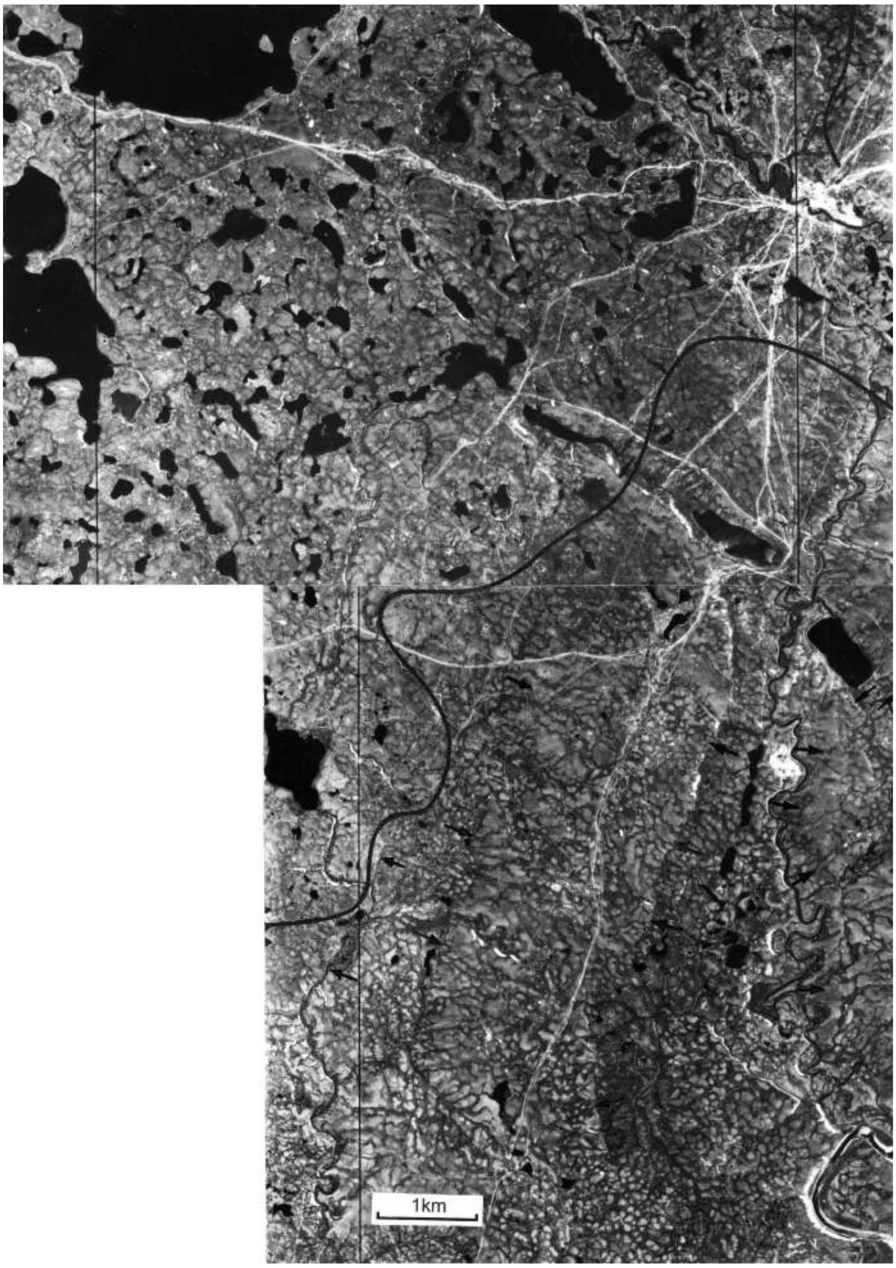

Fig. 8. Aerial photomosaic of the Markhida Line (solid line) south of lake Harbei-To (see location in Fig. 2). Glaciokarst landscape of Harbei type in the upper left part of the picture consists of small morainic hummocks and kames interspersed with deep lakes of intricate configuration; note the hammer-like bays and promontories in the large lakes. The periglacial landscape east of the Markhida Line is shaped by large eroded permafrost polygons best expressed at the bottom of the picture. Proglacial drainage system, indicated by chains of small lakes on flat valley bottoms, consists of two sets of meltwater channels shown by arrows: a couple of older, southwest-oriented marginal channels are cross-cut by a younger radial channel with a beaded system of southeast-oriented elongated lakes. The latter changes in the northwest into a subglacial channel with an uneven floor marked by minor elongated lakes.

Siberia (Astakhov 1979; Astakhov *et al*. 1996). In West Siberia their occurrence hundreds of kilometers upglacier of the ice-sheet margin is connected with the lithology of the deformable glacier bed. In the Pechora Basin the chains of largest glaciotectonic ridges are found much closer to the former ice-sheet margin (Fig. 2).

Ridges consisting of clayey till often have a strongly preferred orientation, coinciding with the axes of hill-hole pairs and the dominant peaks of till fabric. These low longitudinal ridges are interpreted as fluted surfaces. The southwest–northeast oriented valley of the lower Pechora River follows such a streamlined relief, reflecting ice movement from the northeast. Here the normal thickness of the basal till overlying Eemian sand is 2 to 5 m, rarely 10 m (see Sopka and Upper Kuya sections in Mangerud *et al*. 1999). In other areas the apparent till thickness is much greater due to glaciotectonic stacking.

The important geomorphic role of glaciotectonism is inferred by comparing the typical altitudes of the Eemian basement with the altitudes of the present day uplands. The top of the horizontally lying interglacial marine sand is normally found at 40–50 m a.s.l., as can

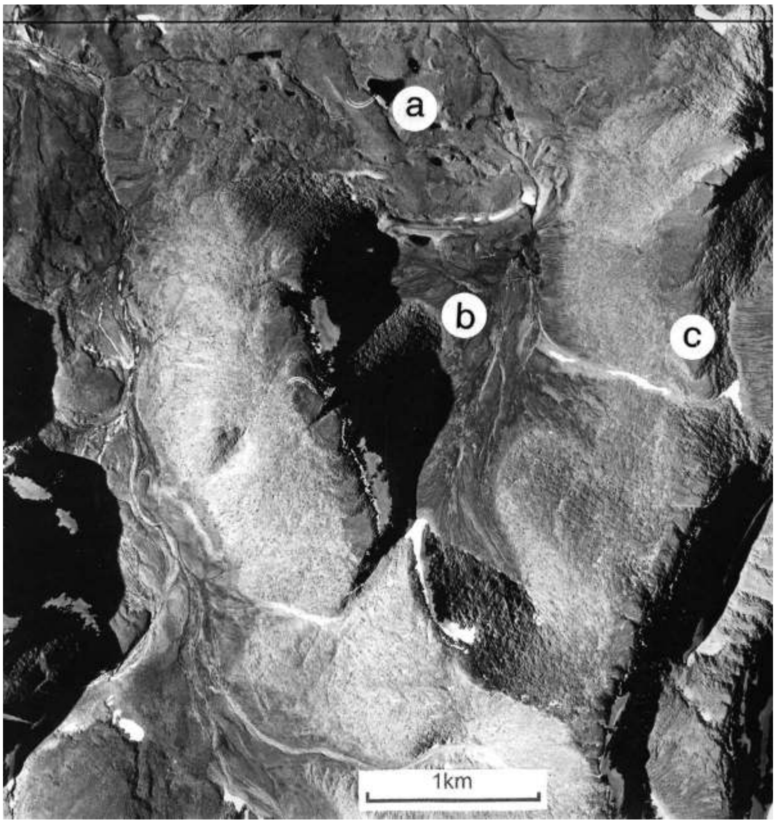

Fig. 9. Aerial photograph of the northwest margin of the Polar Urals (see location in Figs. 2 and 5). (a) – morainic ensemble inserted into a mountain valley from the northwest; a horseshoe-shaped marginal ridge (arrows) is located at 560 m a.s.l.; (b) – flat valley bottom devoid of moraines. (c) – cryoplanation terraces on unglaciated summits at 700–800 m a.s.l.

be seen in sections along Sula River (Fig. 3 see also Mangerud *et al.* 1999) and directly north of the Pechora River mouth. In places where these formations are deeply eroded, as along the Kuya River (7 in Fig. 2) or in the Vastiansky Kon section (8 in Fig. 2), the top of the Eemian marine sediments may be as low as 10–12 m a.s.l. On the other hand, glaciotectonically displaced Eemian sediments may occur up to 100 m a.s.l. in composite ridges.

In the western part, especially along the Pechora valley, there are many 1 to 5 m high knolls, typically a few tens of metres wide, and at places clustered along large till ridges. Arslanov *et al.* (1987) perceived the small knolls as moraines of an Early Holocene ice-sheet advance. However, we found that these knolls consist of soliflucted diamictons, alternating with stratified sand and silt deposited in short-lived thermokarst ponds. The radiocarbon dates from the Markhida section (9 in Fig. 2) indicate that the ponds were formed in the Early Holocene, some 8–10 ka ago, after which they were sediment-filled and topographically inverted during permafrost degradation (Tveranger *et al.* 1995). A similar topographic inversion of perennially frozen terrain has been described in Siberia as the thermokarst cycle (Boitsov 1961; Astakhov 1998a, b). The permafrost origin of the small clayey mounds is evident in aerial photographs of the periglacial zone, where they retain the regular pattern of ice-wedge polygons (Fig. 8).

Apart from the permafrost and fluvial processes, strong wind action is a major factor in shaping the postglacial Markhida landscapes. This is evident from the thick surficial sheets of Late Pleistocene aeolian sand (Mangerud *et al.* 1999) and numerous deflation hollows devoid of vegetation on the sandy tundra. Blowouts must have been ubiquitous in pre-Holocene postglacial landscapes, judging by the wide occurrence of the aeolian mantle on all topographic elements above the flood plain. The sources of the aeolian sand are partly glaciofluvial and glaciolacustrine sandy formations and partly glaciotectonically uplifted thick beds of Eemian marine and fluvial sand. The numerous isolated conical or pyramidal hillocks, 10–30 m high and 30–150 m wide, especially abundant on the left bank of the Pechora River north of 67.5°N, are evidence of previous powerful deflation. Previously, these features were interpreted as large kames (Lavrov *et al.* 1991), even though they are not associated with glaciokarst lakes. The parabolically concave slopes, armoured by angular boulders, split pebbles, and sometimes marine mollusc shells on the surface, and particularly the pointed summits, suggest that these hillocks are the result of wind erosion. Smaller, 1–3 m high armoured cones with

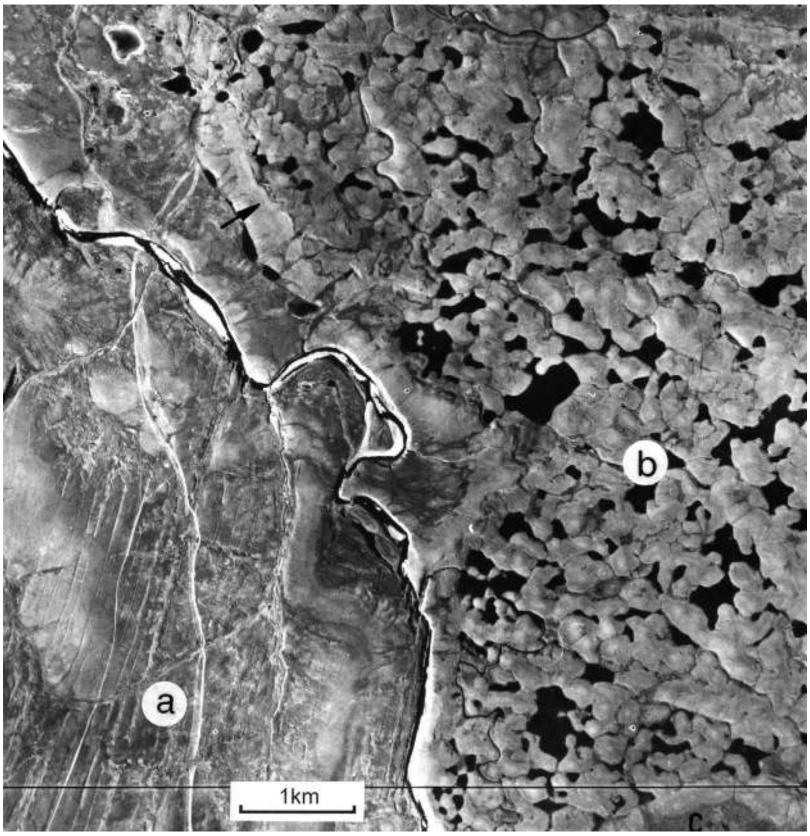

Fig. 10. Aerial photograph of Halmer landscape in the foothills of the Polar Urals (see location in Figs. 2 and 5). (a) – Pemboi Plateau built of gently dipping Permian conglomerates at 200–260 m a.s.l., heavily striated by an older ice flow from the north-northeast. Extremely fresh glaciokarst landscape (b) is separated from the Halmer-Yu River by a narrow ridge (arrow) at 230 m a.s.l. and 30 m above the marginal chain of elongated lakes. A part of a drained intraglacial lake (grey) with a morainic rim (light-grey) is visible in the right bottom corner (c).

ventifacts are common decorations on the sandy surfaces along the large glaciotectonic ridges (Astakhov 1979).

Hummocky morainic landscape of Harbei type

East of the Markhida moraines, beyond the catchment area of the Pechora River, the topography grows perceptibly higher with interfluve plateaus at 170 to 210 m a.s.l. Hummock-and-lake terrains (Fig. 8) are ubiquitous, giving this landscape a younger appearance than the Markhida type. Deep glaciokarst lakes of intricate configuration and of various sizes are characteristic of this landscape, although shallow thermokarst ponds, sometimes surrounded by boggy hasyreis, are also common. The density of small (1–6 hectares) lakes is estimated at around 5–6 per km^2. Wind-eroded depressions are common features, but solifluction flows are perceptibly less frequent than in the Markhida landscape. The type area of this landscape is around lake Harbei-To (Fig. 8). All Harbei-type terrains west of the Urals are underlain by 300–500 m thick permafrost (Yershov 1988) and drain directly into the Barents Sea.

Individual glacial landforms such as morainic ridges and eskers are similar to those of the Markhida type landscape, although their distribution is different. The composite ridges are generally smaller (1–4 km wide) and less prominent (10–15 m high above the background plain). Chains of such ridges are mostly parallel to the southern boundary of the Harbei landscape, outlining north–south oriented lobate basins (Fig. 2). One such chain, probably a retreat moraine, can be traced for 80 km in a southeast–northwest direction between the More-Yu and Chornaya rivers. It separates an area with fresh glaciokarst topography in the northeast from more eroded terrains covered by relict permafrost features in the southwest (Fig. 2).

An unusual ridge, transverse to the strike of the Palaeozoic structures, is visible in satellite images of the southwestern Pai-Hoi (Fig. 2). The ridge, which is 80 km long and up to 200 m wide, is striking north north–south, parallel to the general ice-flow direction. It consists of distorted clayey diamicton up to 40 m thick, interpreted as till material. Judging by its slightly concave form in plan, the ridge is interpreted as a lateral moraine deposited by an ice lobe that advanced southwards across the Pai-Hoi range. Alternatively, it may have an ice-pressed origin from a longitudinal crevasse in a former ice sheet.

Small sandy hummocks resembling kames occur at higher altitudes (150–250 m a.s.l.) near the source of the Adzva River (Vashutkiny lakes), where they mark the southern boundary of the Harbei landscape. More characteristic of the Harbei landscape are isolated oval

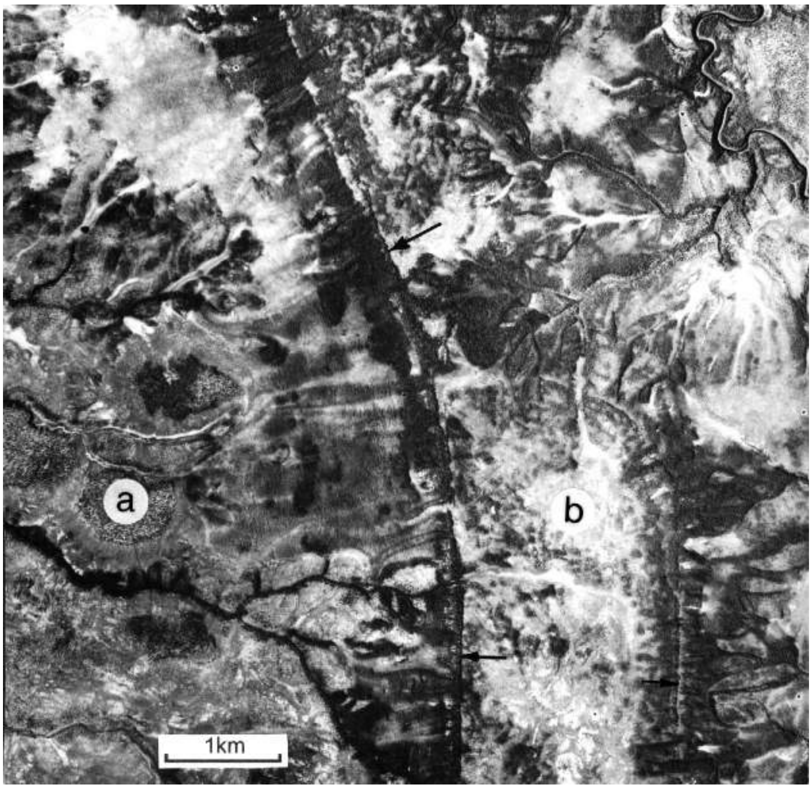

Fig. 11. Aerial photograph of the left bank of Sozva River in the central part of the Pechora Basin (see location in Fig. 2). A well-expressed shoreline, traced along the 100 m isohypse by a narrow sand bar (arrows), separates a flat swampy plain at 85 to 95 m a.s.l. (a) from a promontory of a pre-Weichselian glaciofluvial plateau at 105 to 120 m a.s.l. (b) Note the flat oval mounds (a) on the floor of Lake Komi interpreted as kames by Lavrov (1978) and as inverted thermokarst ponds by the present authors.

or round table-like plateaus 30–40 m high and 30–35 km^2 large, described by Lavrov (1978) as ice-contact features called limnokames. Their surface is flat or concave with a centripetal drainage network. The slopes, 10–20° steep, have sharp knicklines at the foot. The plateaus are composed of laminated silt and fine sand or sometimes of varve-like silty rhythmites with 1000–2000 apparently annual layers. Similar, but smaller (3–5 km^2), forms occur also among the Markhida moraines. In agreement with Lavrov's interpretation we consider them to be glaciolacustrine ice-contact features that originated amongst fields of stagnant glacier ice at the latest stages of the ice-sheet disintegration.

The thickness of glacial sediments overlying the Eemian marine formation may reach 80–120 m on higher plateaus (180–240 m a.s.l.), where they are glaciotectonically stacked together with slabs of interglacial marine sediments (Lavrushin *et al.* 1989). Even the undisturbed uppermost glacial complex is often up to 30–40 m thick due to the very thick (20–30 m) glaciolacustrine rhythmites overlying the basal till.

Within the clayey diamicts there are lenses of massive dirty ice up to 30 m thick described in cores by N. Oberman, who interprets these stratiform bodies as pingo ice formed by water injections (Yershov 1988). We found an exposure of ice at the base of a 30 m high till bluff on the left bank of More-Yu River, 60°E. In this section, steep-dipping minor (5–10 cm thick) bands of clear ice are contained within thicker (1–2 m) layers of ice/diamicton mixture with strongly preferred northeast orientation of numerous angular and wedge-shaped pebbles. The ice is covered by a diamicton of similar composition, with a distinct thaw contact, which is interpreted as meltout till. In appearance and structure this banded ice is similar to the fossil glacier ice described in Siberia (Astakhov & Isayeva 1988; Astakhov *et al.* 1996), but the More-Yu ice contains more clastic material. We believe that such ice bodies, which according to Oberman (Yershov 1988) often make cores of accretion ridges in the southern Pai-Hoi range, are actually remnants of stagnant glacial ice surviving within the thick Pleistocene permafrost. Finds of fossil glacier ice are very important for understanding the fresh-looking hummocky landscapes. Such landscapes are not necessarily signatures of a young ice advance, but can originate in the course of retarded melting of an old glacier ice surviving within thick stable permafrost (Astakhov & Isayeva 1988).

East of the Urals a hummocky landscape of the Harbei type is represented by the Sopkay moraines (Astakhov 1979), which include the northwest–southeast striking Sopkay marginal ridge proper and the hummocky assemblages to the north. In this area, clayey hummocks are interspersed with numerous small sandy kames and stoss-and-lee features on salients of heavily striated Palaeozoic limestones (Fig. 2). Solifluction flows occur on long clayey slopes, but they do

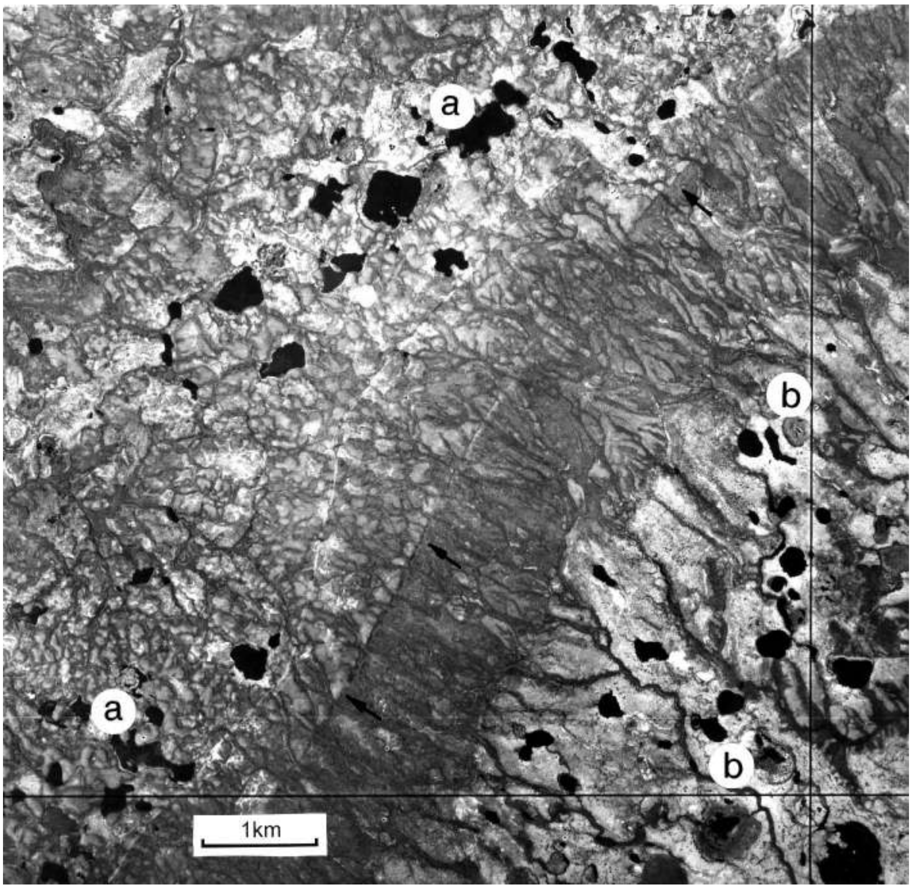

Fig. 12. Aerial photographs of a Lake Komi shoreline north of the Haryaha River inside the Laya-Adzva morainic ridge (location see in Fig. 2). (a) – glaciokarst hummocks-and-lakes of the Markhida landscape type on a plateau at 140–150 m a.s.l.; lakes of intricate shape are often encircled by dry boggy platforms – hasyreis; note also solifluction streams crossing the former lacustrine cliff (arrows) along the 110 m isohypse. (b) – flat swampy bottom of Lake Komi at 90 m a.s.l. with shallow thermokarst lakes and mature hasyreis.

not obscure the glacial hummocky relief. The lithologically diverse substrate is very uneven, which favoured formation of a large intraglacial depression filled with 8 m thick varved clay up to 60 m a.s.l. The Sopkay moraines are slightly modified by three levels of distinct glaciofluvial terraces. The sediments of the last glaciation, according to the borehole data, are normally 20–40 m thick, but reach up to 60–100 m in the main marginal ridge. The solid permafrost is 300 m thick, as in the Harbei landscapes. All directional features, including striae, pebble orientation, stoss-and-lee forms and the strike of the marginal ridge itself, show an ice flow from the north, i.e. parallel to the Ural Mountains. The source of ice must have been located in the Baydarata Estuary of the southwestern Kara Sea (Astakhov 1979).

Hummocky morainic landscape of Halmer type

The most fresh looking hummock-and-lake landscapes occupy the western piedmont of the Polar Urals (Fig. 5 and the dark green colour in Fig. 2). Unlike the Markhida and Harbei moraines, which are underlain mostly by unconsolidated Quaternary and Mesozoic formations, the Halmer moraines rest on a firm basement of Permian and Triassic conglomerates. These moraines are named after the Halmer-Yu River. Spatially, the moraines of Halmer type generally correspond to the landscapes of 'the second phase of the last glaciation' by Kaletskaya (1962). Large glacial lakes are rare here, but the density of minor lakes (1 to 6 hectares) is twice that of the Harbei landscape: around 11 lakes per km^2. The lakes are jammed between small roundish hummocks and have a typical glaciokarst hammer-like form. No shallow thermokarst ponds were noted. The drainage network, represented by short channels connecting the lakes, has apparently just started to develop. There are no deflation hollows, solifluction streams, old polygonal patterns, boggy platforms or other traces of lake regression. In general, the landscape is very similar to the morainic landscape underlain by buried glacial ice along the Yenissei River in Siberia.

The Halmer moraines, outlining a distinct lobate basin with a north–south axis along the upper Kara River, have sharp boundaries with the surrounding bedrock uplands (Figs. 2 & 5). The bedrock frame of the lobe in the east is represented by the frontal escarpment of the Urals with north–south oriented stoss-and-lee features, and in the west by the Pemboi Plateau built of Permian conglomerates which are heavily striated by a preceding ice advance (Fig. 10). West of the Pemboi upland (Fig. 5), beyond the area of Permian and Triassic conglomerates, gentler clayey or sandy hummocks with occasional short eskers gradually change into Harbei type moraines and glaciokarst lakes become less frequent.

The western boundary of the morainic lobe is a 200–250 m wide and 15–20 m high marginal ridge broken into 2 km long arc-shaped segments. It abuts the Pemboi Plateau at 270 m a.s.l. and descends to 218 m a.s.l. at the southern tip of the lobe (Fig. 5), where it merges into a proximal agglomeration of steep (20–25°) hummocks. The latter consist of rounded pebbles in a silt–sand matrix. We have not seen any big boulders there, even as close as 6 km from the front of the Urals. The rounded pebbles of Uralian rock types with predominating quarzite are common for the Permian and Triassic conglomerates underlying the Halmer moraines and exposed at the Pemboi Plateau. Thus, the Perm-Triassic sedimentary rocks along the western foothills are probably the principal source of stones in the Halmer till and not the intrusive and metamorphic rocks of the central Ural Mountains.

In the eastern part of the Halmer morainic lobe, no continuous marginal ridges were observed. Instead, there is a series of horseshoe-shaped morainic ridges apparently shoved from the west into the upper Kara valley at altitudes 240–250 m a.s.l. (Fig. 5). The easternmost ridge is distally fringed by flat proglacial surfaces resembling lacustrine or glaciofluvial terraces. They are surrounded by mountain slopes heavily dissected by ravines ending at the proglacial flats.

A similar pattern is recognized at the northwestern escarpment of the Urals. Here a hummocky landscape intrudes into alpine valleys, where they end up with horseshoe-shaped ridges with convex distal slopes facing up-valley from 280–560 m a.s.l. (Fig. 9). At these altitudes no down-valley morainic loops of local alpine glaciers have been found. Upstream of the south-facing ridges there are only smoothed valley floors, sometimes with dammed elongated lakes. An end moraine ridge across a mountain valley at 560 m a.s.l. (a in Fig. 9) has an asymmetric profile with the steepest (35°) slope along its north-oriented concave side. It is comprised of big boulders in a sand-gravelly matrix. The stones are mostly local Uralian schists and quartzites, but the morainic ridge also contains fragments of limestone that occur *in situ* only at lower altitudes to the north of the ridge. Local down-valley oriented end moraines are only noted higher than 600 m a.s.l. and always very close to the present-day glaciers.

The pattern of the Halmer end moraines together with the provenance of erratics unambiguously shows their origin from a thick inland ice that advanced from the north along the Urals. In the local environment this ice stream could only come from the southwestern Kara Sea shelf, which is confirmed by the pattern of striae and drift ridges on the Pai-Hoi range (Fig. 2) and large erratic blocks transported southwestwards from the Baydarata Estuary coast (Voronov 1951).

Piedmont Moraines from Ural mountain glaciers

Just south of the Halmer moraines, along the Usa River at the foot of the Ural Mountains, there is another morainic assemblage of quite a different orientation. Its axis strikes west–east, pointing to the upper Usa valley, whereas the axis of the Halmer lobe is north–south and parallel to the Urals (Figs. 2 & 5). This morainic apron is outlined by a narrow ice marginal ridge, 20 m high and 40 km long, terminating on a bedrock plateau 180–280 m a.s.l. A wedge of sorted glaciofluvial gravel, looking like a sandur, starts from the western slope of the ridge and descends the Usa River to merge with the Third alluvial terrace (Gesse *et al.* 1963). This terrace must be older than the Second terrace with radiocarbon datings in the range of 24–37 ka BP (Fig. 13 in Mangerud *et al.* 1999).

The ridge is composed of a sand/gravel diamicton with large sub-angular boulders of granite, gneiss, quartzite, schist and other rocks of the central Ural Mountains. These materials are much coarser and of different provenance than those of the Halmer moraines. The provenance of the boulders and the fan-shaped form of the moraines indicate that they were deposited by a piedmont glacier which originated from merged valley glaciers. The hummock-and-lake morphology east of the ridge makes a sharp contrast to the soliflucted and permafrost patterned plateaus to the west of it. In the centre of the lobe basin hummocks are fragmentary and subdued, with solifluction streams over the bedrock salients. Lakes are small and scarce, with a density of about 2 per km^2. The drainage network is much better developed than within the Halmer morainic lobe. A similar apron of local moraines occurs also in the eastern Urals immediately south of the Sopkay moraines (Astakhov 1979). South of the described Usa piedmont lobe only small horseshoe-shaped morainic loops of individual valley glaciers have been mapped along the western front of the Polar Ural Mountains (Gesse *et al.* 1963).

The relations between the piedmont moraine and the Halmer moraines deposited by an ice stream from the Kara Ice Sheet are essential. As appears from the map (Figs. 2 & 5), along the Mal Usa River there is a 20 km long southern continuation of the Halmer morainic lobe. This continuation, directly bordering on the piedmont moraines, is also oriented along the front of the Urals and is fringed by similar marginal ridges, as in the northern Halmer lobe, and was therefore also deposited by a south-flowing ice stream (Fig. 5). The hummock-and-lake landscape within this southern lobe is more subdued compared to the Halmer landscape, with flat bogs, solifluction streams, old polygonal patterns, fragmentary riverine terraces and shallow thermokarst lakes. Judging by the stoss-and-lee features on some limestone salients, the drift within this part of the lobe is thin. Although the peri-Halmer moraines must have been deposited by the same ice stream as the Halmer moraines proper, they are morphologically closer to the more eroded landscapes of Harbei type. Accordingly, the area inside the lobe is mapped as hummocky landscape of Harbei type and shown by a light-green colour in Fig. 2.

In the 2 km wide zone of confluence between the peri-Halmer and the piedmont moraines there are some angular small lakes and a cluster of west–east striking eskers. The end of the north–south striking morainic lobe grows oblate, and, when widening, issues westwards three small (0.5–1 km wide) tongues of hummocky moraines. These are indications of an axial compression of a thin ice lobe moving south which collided with the frontal obstacle of a thicker piedmont glacier. Therefore, the piedmont glacier must have developed either simultaneously with or prior to the ice sheet lobe advancing from the north. This conclusion is confirmed by the pattern of meltwater channels. They start from the fresh moraines of the Halmer lobe to cross its southern continuation and cut farther south into the piedmont moraines before merging with the Usa River valley (Fig. 5). This south-bound drainage system, crossing the present drainage ways at the foot of the mountains, could develop only as ice-walled channels. Therefore, in spite of the morphological difference between the Halmer and the local Ural piedmont moraines, they probably belong to the same glaciation.

Glacial drainage features

Glaciofluvial forms are infrequent north of the Markhida Line, but are common along this boundary, especially when the Markhida Line proceeds across terrains higher than 100 m a.s.l. (Figs. 7, 8, 9). Eskers occur mostly on the Palaeozoic rocks of the Timan Ridge or Uralian piedmont. The longest esker (12 km) has been described at Harius lakes (1 in Fig. 2), where it is a north–south striking flat-topped ridge, 10–15 m high and 100–150 m wide, and is composed of cross-bedded sand with gravel and shell fragments. An occasional swarm of eskers on the soft substrate is located along the upper Shapkina River close to the ice margin (Figs. 2 & 4). Some of them are situated in straight dry valleys with uneven floors and beaded profiles characteristic of channels of subglacial drainage. Beyond the inferred ice-sheet margin the meltwater channels and eskers continue as (often dry) valleys with bevelled (lacking spurs) slopes and graded floors. Small lakes and erosional residuals along the thalweg demonstrate a former major meltwater stream (Fig. 7). The transition of eskers and subglacial channels into proglacial dry valleys, independently of the present-day drainage, is the most reliable morphological combination for delineating the ice-sheet margin on uplands (Figs. 7 & 8).

The direction of glacial drainage indicated by eskers and subglacial channels in many cases does not coincide with the general ice flow from the northeast (Fig. 2). This may reflect a thin marginal ice or a late development of the subglacial drainage network.

Laya-Adzva and Rogovaya moraines

South of the Markhida and Harbei hummocky landscapes there are two spectacular loops of large parallel ridges along the Kolva and Rogovaya rivers (Fig. 2). Both morainic systems are truncated by the Markhida Line and are therefore older. The western double-ridge system, which is 250 km long and up to 20 km wide, was described by Lavrov (1966) as the Laya-Adzva Ridge. The most topographically expressive is the 2–5 km wide inner ridge with a 20–30° steep proximal slope rising 40–70 m above the swampy flatland. The latter is positioned at 90 to 100 m a.s.l. According to coring data the ridge consists mostly of distorted clayey diamicts alternating with blocks of stratified silt and sand (Lavrov 1966). The outer ridge is more subdued. Along the crests of both the outer and inner ridge there are numerous small, parallel ridges showing glaciotectonic compression. Towards their ends the main ridges become narrower and sinuous, although retaining their height. The Laya–Adzva moraine is broken only by a few gaps, including the Kolva River and several deep elongated lakes that are up to 5 km long and 1 km wide. Several long, narrow and deep (more than 15 m) lakes accentuate the sharp knickline of the proximal slope of the inner ridge. The very deep (25–60 m) lakes have linear coastlines, giving the impression of glaciotectonic grabens. They are the only large interfluve reservoirs existing south of 67°N in our study area.

The northeast–southwest trending morainic ridges along Rogovaya River (Fig. 2) are gentler and in many places barely protrude from the surrounding swamps. Still, they are easily recognizable even in high-orbit satellite images (Arkhipov *et al*. 1980). According to local geological surveys the ridges consist of very thick (up to 80 m) diamicton resting on Cretaceous sandstones. In one of the ridges undercut by the Seyda River we observed a clayey till with a strong northeast fabric, some 40 m of apparent thickness. The long Rogovaya morainic loop, parallel to the Urals, is important evidence of a persistent glacier flow from the northeast (the Kara Sea source) and the absence of a Uralian ice dome also prior to the Markhida ice-sheet advance.

Old eroded morainic plateaus

South of the Markhida Line and beyond the Laya-Adzva and Rogovaya morainic loops the landscape is deeply eroded by a well-developed arborescent drainage system. Rare morainic ridges that survived on the gently rolling plateaus at 150–250 m a.s.l. are degraded by slope processes. The most striking feature is the total lack of interfluve lakes; small ponds occur mostly within extensive flat bogs. Wide river valleys contain two or three alluvial terraces above the present-day flood plain. The tightly spaced valley network occurs on a variety of substrata, including folded Palaeozoic rocks of the Timan Ridge. Modern, 10–100 m thick surficial permafrost occurs in the tundra of the Usa River catchment area. The rest of the area is covered by boreal forests, where permafrost only exists as a Pleistocene relict layer buried under more than 100 m of thawed rocks (Yershov 1988).

Most Russian glacial geologists consider the upper till of these fluvially dissected plateaus to correlate with the Moscow (Saalian) glaciation. However, Arslanov *et al*. (1987) and Grosswald (1993) draw their Weichselian ice limit across this area (Fig. 1). Our observations show that the upper till of the southern lakeless plateau along the Sula river is overlapped either by Eemian marine formations (sections 3 and 5 in Fig. 2 and 12, 14, 15 and 21 in Fig. 3) or by Saalian sand dunes (section 7 in loc. 4, Fig. 3) (Mangerud *et al.* 1999). A pre-Eemian age of the upper till of the southern dissected plateaus is also supported by several non-finite radiocarbon dates from Pechora River terraces incised into this landscape.

Along the Urals the eroded moraines are covered by a continuous mantle of loess-like silts. West of the Rogovaya River the silts are thin, discontinuous and often replaced by sheets or dunes of eolian sand (see below), or solifluction aprons over valley slopes. Soliflucted diamictons up to 7–10 m thick, are often registered in core profiles across the Pechora valley (Yudkevich & Simonov 1976).

Periglacial features

A much wider distribution of permafrost and eolian features compared to the present-day processes indicate earlier periglacial environments in the study area. North of the Markhida Line, we have described Late Pleistocene solifluction sheets, inverted thermokarst ponds, deflation residuals, sand dunes and cover sand. South of the Markhida Line such features are even more diverse, widespread and well developed.

The best manifestation of former thicker and colder permafrost is the polygonal relief, very characteristic of the northeastern part of the old morainic landscape (Fig. 2), where surficial permafrost was extinct for most of the Holocene and reappeared some 2–3 ka ago (Yershov 1988). This area is dotted by flat mounds, 1–3 m high, outlined by a rectangular net of troughs marking former ice wedges. The resultant polygonal system is reflected in a coarse-grained, shagreened pattern on the aerial photographs (Fig. 8). The present-day permafrost table along the troughs is often 5–10 m deep, i.e. disconnected with the present layer of seasonal freezing. The polygons, from 10–15 m to 300–400 m, and sometimes up to 1000 m across, are incompatible with the present-day permafrost and demand a much colder and more continental climate of North Siberian type (Popov 1962). In the present-day climate only small ice wedges grow near the coast to the north (Yershov 1988).

The former ice-wedge cracks are mostly filled with loess-like silts or soliflucted diamictons. In the foothills close to the Urals, the surficial diamicton is often coarse and in places attains 3–4 m of thickness, and has been interpreted as a till (Kaletskaya 1962). We have observed that in some places this diamicton is blanketing river terraces, together with a mantle of loess-like silts. We therefore conclude that it is a solifluction deposit. This is strongly supported by its spatial connection to the pattern of relict permafrost polygons (Popov 1962).

Near the Urals, the Markhida Line forms a very distinct northern limit for the large relict polygons. They are absent on the Halmer moraines. However, west of 59°E wide tracts of relict polygons intrude into the Harbei morainic landscape from the south, mostly along wide meltwater channels (Fig. 2). Westwards, in the old eroded morainic landscape, the polygons are being obliterated by solifluction and eolian sediments and are therefore not easily recognizable in aerial photographs, especially in boreal forests.

Radiocarbon dates from organic infills in large ice-wedge casts along Shapkina and Kuya rivers yielded ages of about 11.5 and 12.3 ka (Mangerud *et al.* 1999), suggesting that the thick Pleistocene permafrost started to degrade during the Allerød–Bølling interstadials. The polygonal pattern in the Markhida and Harbei landscapes indicate that many places north of the Markhida Line were free of stagnant glacier ice in the Late Weichselian.

Other features interpreted by us as the result of former permafrost are regularly spaced sandy mounds on the unfrozen low flatlands (40–60 m a.s.l.) and on the Pechora River terraces to the south of Shapkina River (Fig. 2). These oval or pancake-shaped mounds are normally several hundred metres across, less than 10 m high and weakly elongated in a west–northwestwards direction. They are composed of fine laminated sands and silts and were interpreted by Lavrov (1978) as supraglacial kames. Some of the smaller sandy mounds can be seen just below the shoreline of a former proglacial lake (Lake Komi) described below (Fig. 11). We interpret these mounds to reflect former thermokarst lakes that developed on perennially frozen ground at the floor of the former ice-dammed lake (Fig. 12). After degradation of the permafrost and general sagging of the surrounding icy terrain, thawed sediments deposited in thermokarst ponds appeared inverted into smooth oval mounds (Boitsov 1961; Astakhov, 1988a, b).

The mantle of loess-like silts, occurring in the same area as the large ice-wedge casts, is up to 3 m thick along the Uralian foothills (Popov 1962) and reaches a thickness of 8 m south of our study area (Kuznetsova

1971). Loess-like silts also occur sporadically in the Markhida type landscape, but have not been found in the Harbei and Halmer landscapes.

Unlike loess-like silt, aeolian sand is widespread in the study area and is commonly recognized in aerial photographs by numerous light-coloured spots of present-day blowouts stripped of vegetation. The sand occurs in the form of dunes and also as mantle-like cover sand. It is normally fine or medium sand with pitted or sometimes varnished surfaces on quartz grains. The indistinct diagonally bedded dunes typically overlie or laterally replace sheets of cover sand with more distinct, fine wavy stratification. The blanket of cover sand, typically 1–6 m thick, occurs on many valley slopes and terraces above the present-day flood plain. In several localities along the Pechora River: Akis (11 in Fig. 2), Byzovaya (21 in Fig. 2), Nyasha-Bozh and Denisovka villages, cover sand grades into pale-brown, crudely laminated loess-like silt. Although in places the aeolian sand may be more than 15 m thick (Mangerud *et al.* 1999), its patchy distribution cannot be shown in the small-scale map. Thick accumulations of aeolian sand occur most frequently in the wide valleys of the Pechora and Usa rivers and on the lowlands along the Barents Sea.

Traces of ice-dammed lake Komi

South of the Markhida Line there are tens of kilometres of wide swampy flatlands along the major rivers. They are separated from higher morainic plateaus by long gentle slopes. In many places along the slopes there are very distinct and laterally persistent breaks along the 100 m (90–100 m) isohypse, looking like erosion notches or strandlines, sometimes accompanied by narrow ridges. We call such linear features knicklines (Fig. 2) to designate the sharp change of gradient across these lines (Figs. 11 & 12) without genetic implications. Many sections slightly below the knicklines have been studied in gravel pits and road cuts near Ust-Tsilma (12 in Fig. 2) and along the Kolva river (locations 13 to 18, Fig. 2). All of them show a sequence of laminated and rippled beach sand coarsening up into fine, well-sorted gravels without traces of marine life or other organics (Mangerud *et al.* 1999). The thickest beach gravel (up to 17 m) with horizons of ice-wedge casts at the base, in the middle and on the top of the sequence is exposed in a ravine at the upstream end of Byzovaya Village, 90 m a.s.l. (21 in Fig. 2). The thinnest gravel (1–1.5 m) has been observed in the northernmost pit on Haryaha River, 110 m a.s.l. (16 in Fig. 2).

From these observations and from the fact that the knicklines never proceed to the sea coast (Fig. 2) we infer that they are shorelines of a large ice-dammed lake. The flatlands below the old shoreline, painted light blue in Fig. 2 are interpreted as the bottom of the lake. We propose the name Lake Komi for this ancient reservoir, which according to our reconstruction occupied all lowlands of the Komi Republic south of the Markhida Line. Fine-grained deep-water rhythmites, observed only in a few sections (Mangerud *et al.* 1999), are mostly known from geotechnical cores taken upstream along Kolva River. Commonly, the floor of Lake Komi, when exposed at altitudes of 70–90 m a.s.l., is covered by only 1–5 m thick Holocene peat or a thin mantle of aeolian sand and loess-like silt on top of an extremely flat till surface. Below the 70 m level the lake bottom is replaced by two younger fluvial terraces.

In places the strandline is highlighted by a narrow sand bar (Fig. 11) or a low cliff in consolidated or permeable sediments (Fig. 12). Along many till uplands erosional notches are missing, probably being obliterated by solifluction on poorly drained clayey slopes. Also along bedrock uplands, such as upstream of the Usa River catchment area (Fig. 2), the shorelines are not always morphologically expressed. The lake was probably too short-lived to produce cliffs in hard Palaeozoic rocks.

We have traced the Lake Komi shorelines in aerial photographs westwards across the Timan Ridge (Fig. 2). A 40 km long and 20 m wide sand bar along the 110 m isohypse borders a flat sandy embayment of the upper part of the Tsilma River valley. Farther to the west the embayment narrows into a 2 km wide passage with a high moor bog at 113 m a.s.l. Across this watershed knicklines can be followed for 4–6 km as graded cliffs covered by solifluction sheets along the 120 m isohypse. To the west of the Timan Ridge and south of Pyoza river, a knickline, seen on high-resolution satellite images, persists along the 110 m isohypse, in places underlined by sand bars. We therefore conclude that Lake Komi continued across the described watershed to coalesce with a similar reservoir in the Mezen River catchment area.

East of the Pechora River a very pronounced knickline along the 100 m isohypse can be traced continuously from the village of Ust-Tsilma (12 in Fig. 2) to Shapkina River. In places it outlines some narrow headlands of older till and outwash massifs protruding into the proglacial lake (Fig. 11). North of Shapkina River a knickline at 110 m a.s.l. intrudes into the Markhida landscape by embayments with outliers of morainic plateaus (Figs. 2 & 4). Farther northwards the former shorelines disappear, being replaced by hummocky terrains with lower altitudes.

The strandlines are well developed inside the Laya-Adzva moraines (Fig. 2). They penetrate far northwards along Kolva River, cutting into the glaciokarst landscape of the Markhida type (Fig. 12). Varved glaciolacustrine silt and clay covering the till at the site Hongurei (6 in Fig. 2 Mangerud *et al.* 1999) is another indication that Lake Komi inundated the Markhida moraines during the deglaciation.

In the east, the only place where the bottom of Lake

Komi contacts the Markhida Line is the upper stretch of Rogovaya River (Fig. 2). Here, the Harbei moraines at 130–150 m a.s.l. are fringed by a more than 30 m thick crescentic wedge of glaciolacustrine sand covered by glaciofluvial gravel interpreted as a sandur. The distal edge of this sandur merges with the lake level at 100 m a.s.l. Along the upper Adzva River, incised into a higher plateau, a gravelly infill of a meltwater channel seems to cut into the lake bottom at 80–90 m a.s.l. This means that meltwater discharge from the ice sheet in the north was maintained also after Lake Komi was drained. In most cases, however, floors of the proglacial meltwater channels do not descend below the bottom of Lake Komi, which is normally cut by only present-day rivers.

The distribution of the shorelines, and the described morphological relationships with the Markhida Line, strongly indicate that Lake Komi was dammed by the ice sheet that formed the Markhida Line.

Other flat terrains

Isolated flatlands

Swampy flatlands occur also as patches confined within the hummocky landscapes (dark blue in Fig. 2). They are characterized by numerous roundish and shallow thermokarst lakes, and often surrounded by peaty platforms of hasyreis with fine reticulate pattern of small present-day frost-crack polygons. They are positioned at altitudes from 30 m up to 170–180 m a.s.l., the latter being located in the western Timan Ridge (1 in Fig. 2). The underlying sediments are laminated sand and silt or, as in the case of the former lake within the Sopkay moraines, varved clays (Astakhov 1979). The diverse altitudes and lack of spillways suggest that these isolated flatlands are floors of late and post-glacial lakes which developed after the final stagnation of the ice sheet. Radiocarbon dates from the limnic sequences show that some of these lakes still existed 8–9 ka ago (Arslanov *et al.* 1987; Lavrov & Potapenko 1989). We have obtained three radiocarbon dates in the range 13–12 ka BP from silty rhythmites of a drained lake on the west–east stretch of the More-Yu river.

Coastal lowland

Along the Barents Sea coastline there is a flat or seawards inclined lowland at altitudes of from 10 to 40 m a.s.l. (yellow in Fig. 2). It is perennially frozen with many shallow thermokarst ponds but no morainic hummocks or deep glaciokarst lakes have been noticed. The coastal flats have traditionally been mapped as marine terraces (Krasnov 1971; Arslanov *et al.* 1987). Apart from the present-day marshes below 5 m a.s.l., we found no coastal cliffs, beach bars or other morphological evidence of higher sea level, except Eemian marine sediments. Moreover, the flatland gradually climbs inlands and intrudes the morainic landscape in the form of long sandy tongues which in places rise to 60 m a.s.l. On the western shore of the Pechora Estuary there are some narrow, 10–40 m high, laterally not persistent erosional steps carved in the Eemian marine formation. This staircase is thought to be a product of fluvial erosion during the drainage of Lake Komi. Alternatively, they may be cryoplanation terraces formed by frost action and modified by wind erosion during the late glacial time.

Our sedimentological investigations on the Timan Beach (2 in Fig. 2) have found only aeolian sands underlain by finely laminated limnic sand with wisps of soliflucted diamictons (Mangerud *et al.* 1999). In the east the flatland is built of a thin till overlying an interglacial marine formation and covered by sand (Lavrushin *et al.* 1989). Along the lower stretch of the More-Yu River thick eolian sands are directly underlain by till or by interglacial marine sediments.

In summary, the coastal lowland probably originated as a system of glacially eroded depressions subsequently modified by subaerial erosion and deposition during a cold and arid period of low sea level. The present-day configuration of the Barents Sea coastline, as manifested in the system of large and shallow estuaries (Fig. 2), is a product of the Holocene transgression encroaching onto this perennially frozen lowland.

Discussion

Contrast between the hummocky landscapes in the north and older morainic plateaus in the south

The main morphological boundary of the study area, the Markhida Line, is pronounced west of the Pechora valley, where the hummocky Markhida moraines directly border on the old dissected lakeless morainic plateaus. In this area the Markhida Line is marked by a continuous chain of ice-pushed composite ridges (Figs. 2 & 3). There is also a very sharp geomorphological boundary close to the Urals, where the rugged glaciokarst Harbei and Halmer landscapes are abruptly replaced by a gently rolling lakeless tundra with many relict permafrost features (Figs. 8 & 10).

The Laya-Adzva and Rogovaya ridges represent a morphological transition between these two extreme cases. The Markhida Line is less distinct also in the area between these ridges and the Lower Pechora valley, where it proceeds across the low-lying terrains that were flooded by the ice dammed Lake Komi (Figs. 4, 6 and 12). The sharpness of the Markhida Line is also diminished in the central part of the region by the relict permafrost features, penetrating northwards into the hummocky landscape between the Shapkina and More-Yu rivers (Fig. 2).

Lateral correlation of the Markhida, Harbei and Halmer morainic landscapes

We divided the hummock-and-lake assemblages north of the Markhida Line into three types according to the degree of their modification by postglacial processes. In this respect the Halmer moraines in the western foothills of the Urals are the freshest and the Markhida type the most eroded. Glacial landscapes of the Harbei type are of intermediate preservation. The boundaries between the three landscape types are gradual, and unlike the Markhida Line they do not coincide with pronounced marginal features. It is noteworthy that the morphological difference between the three landscape types is better expressed close to the Markhida Line, while northwards the boundaries between the hummocky landscapes are more diffuse.

Theoretically, the three types of morainic landscapes could be attributed to ice sheet limits of different age. We have found little geomorphic and no stratigraphic evidence to support such an interpretation. The lack of marginal features between the hummocky landscapes, the general conformity of the entire pattern of ice-pushed ridges to the west–east trend of the Markhida Line, the same succession of sedimentary formations on top of and beneath the basal till indicate that at least the Markhida and Harbei moraines were left by the same ice sheet. Furthermore, the meltwater channels from both the Markhida and Harbei moraines seem to have emptied into one single proglacial lake, i.e. Lake Komi.

The Halmer landscape is a special case because of its higher position on the Uralian piedmont and the lack of overlying sediments. The boundary with the Harbei landscape is very indistinct and not associated with end moraines or other indications of a former glacier limit. The pattern of the ice marginal features along the southern boundary of this landscape (Figs. 5 & 9) indicates an ice-flow direction from the Kara Sea coast towards the Urals. From a glaciological point of view it is nearly impossible to visualize a former ice cap localized within the Halmer landscape. Such a small ice cap could not have maintained the uphill ice flow up to 560 m a.s.l. in the Uralian valleys. As discussed below, we assume that the fresh appearance of the Halmer landscape is related to the permafrost conditions and late melting of buried glacier ice.

Permafrost distribution and evolution of the hummocky landscapes

The geographical distribution of the Markhida, Harbei and Halmer landscape types roughly corresponds with the zonality of present-day permafrost; the freshness of glaciokarst topography increasing with the growth of permafrost thickness towards the east. The Harbei moraines, underlain by the thickest and oldest permafrost, contain numerous bodies of massive ice which, at least partly, are of glacial origin. We therefore relate the west–east trend in morphological maturity to the melting history of the buried glacier ice and Pleistocene permafrost. Presumably, buried glacier ice in the western sector, where climate was milder and the permafrost thinner, melted away earlier than in the east. The Markhida moraines, especially the higher ice-pushed ridges, were therefore first to be eroded by fluvial processes, solifluction and wind action. When the Pleistocene ice-wedge polygons developed on the Markida and partly Harbei moraines, large fields of stagnant glacier ice probably still existed in the area where the Halmer landscape developed later. The purely glaciokarst landscape of the Halmer moraines, devoid of postglacial fluvial features, was probably formed only after the start of the Holocene climatic amelioration. This accounts for the lack of Pleistocene frost polygons and aeolian cover sand. Such a retarded deglaciation was previously described from the Yenissei River in Siberia, where fresh morainic landscapes are developed over Early Weichselian basal till with large stratified bodies of fossil glacier ice (Astakhov & Isayeva 1988). This implies that the evolution and morphological appearance of Arctic glacial landscapes is to a large extent predetermined by permafrost conditions, which are dependent on the west–east climatic gradient across the Russian mainland.

Margin of the last Barents and Kara ice sheets across the Russian mainland

The Markhida Line represents the southern boundary of the last Kara and Barents ice sheets. Along its western stretch (up to Kolva River in the east) the Markhida Line is nearly identical to the ice margin suggested by Yakovlev (1956) for his Early Weichselian glaciation. Yet, contrary to Yakovlev's and many other interpretations (cf. also Kaletskaya 1962; Krasnov 1971; Lavrov 1977; Arslanov *et al.* 1987; Biryukov *et al.* 1988; Faustova & Velichko 1992, Velichko *et al.* 1997), our mapping results testify to the absence of a contemporary ice cap upon the Urals. This is evident from the latitudinal trend of the ice margin across the Palaeozoic salients (Fig. 2). The pattern of the ice-pushed ridges, striae and till fabrics points to the Kara Sea as the main source of moving ice for all mapped glacial stages. During the last ice advance from the Kara shelf, alpine glaciers of the Urals were too small to obstruct the uphill ice flow of the Kara Ice Sheet.

Glaciological interpretations of the available ice-flow signatures indicate that the thicker ice of the last glaciation was localized in the western Kara Sea and must have inundated the Yamal Peninsula (Tveranger *et al.* 1999). From the pattern of morainic lobes west of the Pechora valley (Fig. 2) we also infer that another ice dome at the same time existed in the Barents Sea and that the Barents and Kara ice sheets merged in the western part of the study area. This is confirmed by the shorelines of Lake Komi, which are traced approximately at the same level from the Mezen River catchment to the Urals. This in turn implies a total blockade of the northbound drainage of the Russian European mainland (Fig. 2). Taking into account the uppermost till of Kolguyev Island, which was apparently deposited by an ice flow from the northeast (Baranovskaya *et al*. 1986), the line of confluence of the Barents and Kara ice sheets can be drawn immediately to the west of this island (Fig. 2).

Minimum dates of deglaciation within the Markhida type landscape suggest that the last ice sheet of this area melted away before 45 ka ago (Mangerud *et al.* 1999). This conclusion is supported by radiocarbon dates of sediments overlying the latest Kara till on the western coast of the Yamal Peninsula. Here, two series of AMS dates have been obtained from eolian silts (12.2–16.4 ka, locality 24 in Fig. 2) and from thick peats (28.4 to 32.7 ka, locality 25 in Fig. 2) (Gataullin & Forman 1997; Gataullin *et al*. 1998).

Lake Komi and its relations with the Kara and Barents ice sheets

The photointerpretation has shown that the strandlines of Lake Komi are incised into the hummocky landscape of the Markhida type (Figs. 4 & 12), terminating not far north of the Markhida Line (Fig. 2). From hence we infer that Lake Komi, which was dammed by the same ice sheet that left the Markhida moraines, also existed during early stages of deglaciation. This is supported by many meltwater channels and a sandur merging with the 100 m level of Lake Komi in front of the Markhida and Harbei moraines. Yet, some other meltwater channels cutting this level indicate that Lake Komi was drained well before the final decay of the fields of stagnant glacier ice.

Therefore, sediments overlying the former lake bottom are more likely to yield older minimum dates for the last glaciation than sediments atop the hummocky moraines, where buried glacier ice melted over a considerable span of time. Two alluvial terraces, which along with the flood plain descend downstream the major present rivers, postdate Lake Komi. A series of radiocarbon dates from the fluvial terraces, incised into the floor of Lake Komi, and from Palaeolithic sites along the Pechora, Usa and Adzva rivers (locations 19–23 in Fig. 2), show that the present-day northbound drainage was restored well before 37 ka ago (Mangerud *et al.* 1999). An Early/Middle Weichselian age of the ice-dammed lake and thereby of the last glacier that terminated along the Markhida Line is also indicated by luminescence dates ranging from 76 to 93 ka obtained on beach sediments of Lake Komi (Mangerud *et al.* 1999).

So far we have not been able to identify spillways from Lake Komi. One possibility we have not explored is that the lake water overflowed the southern water divide of the Pechora Basin. In the upper Pechora reaches there are morphologically indistinct through valleys at 130–140 m a.s.l., directed south to the Caspian Basin. These spillways, which are higher than shorelines of Lake Komi were earlier related to the Middle Pleistocene glaciations (Krasnov 1971). However, considering that the present surface of this threshold is underlain by thick Quaternary sediments, a possibility of Lake Komi draining southwards cannot be completely ruled out. Another possibility is a spillway into the Barents Sea, perhaps across the Kola Peninsula. If true, this northwestern drainage path would imply an ice-free corridor between the Barents and Scandinavian ice sheets during the Early/Middle Weichselian time. The third, and most likely alternative, is a southwestward drainage via the Onega River basin.

Relations between the Kara Ice Sheet and the Uralian glaciers

Our investigations show that latest glacial ice flowed from the Kara Sea shelf into the Uralian mountain valleys, where it deposited end moraines facing upslope. The only traces of former alpine glaciers in the valleys inundated by the Kara Ice Sheet are coarse tills with blocks of central Uralian rocks composing the upslope oriented end moraines. However, immediately to the south of the ice sheet boundary we identified moraines of a local piedmont glacier in the Bol Usa valley that seems to have existed simultaneously with the adjacent ice-sheet lobe (Fig. 5). This pattern indicates that alpine glaciers developed before the culmination of the last ice sheet, and that end moraines deposited in front of local glaciers in the areas to the north were assimilated by the much thicker Kara Ice Sheet lobes. From the pattern of meltwater channels, starting at the ice-sheet margin and directed southwards, we infer that the piedmont glacier became stagnant before the final melting of the ice sheet lobe. The alpine valleys bear no traces of younger local glaciers overriding the Kara Ice Sheet moraines. This indicates that the local glaciation in the Polar Urals was even more restricted during the Late Weichselian than during the Early/Middle Weichselian.

Another older Weichselian glaciation?

A major remaining uncertainty concerns the huge loops of the Laya-Adzva and Rogovaya morainic systems. Being cut by the Markhida Line and by the shorelines of Lake Komi (Fig. 2), they must be older than both these formations. However, the postglacial morphological modification of the Markhida moraines is not much different from that of the Laya-Adzva and Rogovaya ridges. We therefore find it likely that the Laya-Adzva and Rogovaya moraines belong to a Weichselian glacial stage preceding the ice-sheet advance that dammed Lake Komi.

Conclusions

Three morphologically different types of hummocky morainic landscapes have been photogeologically mapped as a continuous west–east striking belt across the European part of the Russian mainland. Their collective southern boundary, called the Markhida Line, marks the main stationary position of the last Barents and Kara ice-sheet margin. It is mainly confined to the area north of 67.5°N, 100–200 km south of the present coastline. It is dated to be of Middle or Early Weichselian age (Mangerud *et al*. 1999).

The morphological diversity of the glacial landscapes north of the Markhida Line does not reflect different ice advances, but is attributed to different melting histories of stagnant glacier ice and Pleistocene permafrost in the west and east of the region.

The principal Early/Middle Weichselian ice-sheet domes were situated on the shelves of the shallow Kara and Barents seas. Traces of only small alpine glaciers, forming piedmont aprons south of the inland ice margin, have been found in the Urals.

Shorelines of a large ice-dammed lake, called Lake Komi, have been mapped south of the Markhida Line at altitudes between 90 and 110 m a.s.l. The former lake is traced across the Russian mainland, from the Urals to the Mezen River basin, indicating a continuous front of coalesced Kara and Barents ice sheets.

The geomorphic position of radiocarbon-dated river terraces in relation to the bottom of Lake Komi indicates an Early/Middle Weichselian age for this ice-dammed lake and for the ice-sheet margin along the Markhida Line. The Laya-Adzva and Rogovaja morainic loops that protrude 100–150 km south of the Markhida Line were probably deposited by an older Weichselian ice-sheet advance, prior to the formation of the Lake Komi shorelines.

Acknowledgements. – The investigations, which were carried out in the period 1993–98, were funded by the Research Council of Norway, and from 1998 this research was also supported by the EC *Environment and Climate Research Programme* (*Contract no. ENV4-CT97-0563, Climate and Natural Hazards*). Additional financial support was provided by the University of Bergen. This paper is a contribution to the Russian–Norwegian research project *PECHORA (Paleo Environment and Climate History of the Russian Arctic)* and to the European Community Project: *Ice Sheets and Climate in the Eurasian Arctic at the Last Glacial Maximum (Eurasian Ice Sheets)*. Both projects form part of the European Science Foundation Research Programme *QUEEN (Quaternary Environments of the Eurasian North)*. Several Russian and some Norwegian friends assisted during the fieldwork. We appreciate the criticism of Steve Forman, which helped to improve the text. To all these persons and institutions we offer our sincere thanks.

References

Arkhipov, S. A., Astakhov, V. I., Volkov, I. A., Volkova, V. S. & Panychev, V. A. 1980: *Palaeogeography of the West Siberian Plain at the Late Zyryanka glaciation maximum*. 109 pp. Nauka, Novosibirsk (in Russian).

Arslanov, H. A., Lavrov, A. S., Potapenko, L. M., Tertychnaya, T. V. & Chernov, S. B. 1987: New data on geochronology and paleogeography of the Late Pleistocene and Early Holocene in the northern Pechora Lowland. *Novye dannye po geokhronologii chetvertichnogo period*, 101–111. Nauka, Moscow (in Russian).

Astakhov, V. I. 1979: New data on the latest activity of Kara shelf glaciers in West Siberia. *IGCP Project 73/1/24, Quaternary Glaciations in the Northern Hemisphere*, Rpt. 5, 22–31, Prague.

Astakhov, V. I. 1992: The last glaciation in West Siberia. *Sveriges Geologiska Undersökning*, Ser. Ca *81*, 21–30.

Astakhov, V. I. 1997: Late glacial events in the central Russian Arctic. *Quaternary International 41/42*, 17–25.

Astakhov, V. I. 1998a: The last Kara Sea ice sheet: terrestrial constraints on its age. *Quaternary International 45/46*, 19–28.

Astakhov, V. I. 1998b: Inversion relief as indicator of paleocryological environments. *Geomorfologia 4* (in Russian, English summary).

Astakhov, V. I. & Isayeva, L. L. 1988: The 'Ice Hill': an example of 'retarded deglaciation' in Siberia. *Quaternary Science Reviews 7*, 29–40.

Astakhov, V. I., Kaplyanskaya, F. A. & Tarnogradsky, V. D. 1996: Pleistocene permafrost of West Siberia as a deformable glacier bed. *Permafrost and Periglacial Processes 7*, 165–191.

Baranovskaya, O. F., Grigoryev, M. N. & Malyasova, Ye. S. 1986: The stratigraphy of the Late Cenozoic of the Kolguyev Island. *In* Zarkhidze V. S. & Kulakov Yu. N. (eds.): *Kainozoi shelfa i ostrovov Sovietskoi Arktiki*, 83–89. Sevmorgeologia, Leningrad (in Russian).

Biryukov, V. V., Faustova, M. A., Kaplin, P. A., Pavlidis, Y. A., Romanova, E. A. & Velichko, A. A. 1988: The palaeogeography of Arctic Shelf and coastal zone of Eurasia at the time of the last glaciation. *Palaeogeography, Palaeoclimatology, Palaeoecology 68*, 117–125.

Boitsov, M. N. 1961: On development of topography in the conditions of subterraneous glaciation. *Trudy VSEGEI*, ser. nov. *64*, 27–36 (in Russian).

Faustova, M. A. & Velichko, A. A. 1992: Dynamics of the last glaciation in northern Eurasia. *Sveriges Geologiska Undersökning*, Ser. Ca *81*, 113–118.

Gataullin, V. & Forman, S. 1997: Stratigraphy of Upper Quaternary sediments in western Yamal as a key for last glacial paleogeography reconstructions. *International Conference: The Problems of Earth Cryosphere*, 248–249. Pushchino. ISBN 5-201-14333-4.

Gataullin, V., Forman, S., Ingolfsson, O. & Manley, W. 1998: New data on stratigraphy of Quaternary deposits on the western coast of the Yamal Peninsula. *All-Russia Conference: Main Results of Quaternary Studies and Prospects of Research in XXI Century*, 19–21. VSEGEI, St. Petersburg (in Russian).

Gesse, V. N., Granovich, I. B. & Savelyev, A. A. 1963: On the mode of glaciation of the Polar Urals in the Late Pleistocene and Holocene. *In* Popov, AI. & Afanasyev BL. (eds.): *Kainozoisky pokrov Bolshezemelskoi Tundry*, 105–110. Moscow University (in Russian).

Grosswald, M. G. 1980: Late Weichselian ice sheets of Northern Eurasia. *Quaternary Research 13*, 1–32.

Grosswald, M. G. 1993: Extent and melting history of the Late Weichselian ice sheet, the Barents–Kara continental margin. *In* Peltier W. R (ed.): *Ice in the Climate System*, 1–20. NATO ASI series I: Global Environmental Change. Springer Verlag, Berlin.

Grosswald, M. G. 1994: The drumlin fields of the Novaya Zemlya–Urals region and their relation to the Kara ice dispersal center. *Geomorfologia 1*, 40–53 (in Russian, English summary).

Kaletskaya, M. S. 1962: Palaeogeographic peculiarities of the last ice sheet in the extreme north-east of the European USSR. *Materialy po geologii i poleznym iskopayemym Severo-Vostoka Yevropeiskoi chasti SSSR*, 2, 60–66. Moscow (in Russian).

Krasnov, I. I. (ed.) 1971: Map of the Quaternary of the European

USSR and adjacent territories, scale 1:1 500 000, VSEGEI, Leningrad, 16 sheets.

Kuznetsova, L. A. 1971: *The Pleistocene of the Pechora Peri–Urals (Pleistotsen Pechorskogo Priuralya)*. Kazan University, 123 pp. (in Russian).

Lavrov, A. S. 1966: New data on limits of the Boreal Transgression and Kalinin ice sheet in catchments of the Pechora and Vychegda. *In* Grichuk V. P., Ivanova I. K. Kind N. V. & Ravsky E. I.(eds.): *Verkhny pleistotsen*, 112–120. Nauka, Moscow (in Russian).

Lavrov, A. S. 1977: Kola-Mezen, Barents Sea–Pechora, Novaya Zemlya – Kolva ice streams. *In* Chebotareva, N. S. (ed.): *The Structure and Dynamics of the Last Ice Sheet of Europe (Struktura i dinamika poslednego lednikovogo pokrova Evropy)*, 83–100. Nauka, Moscow (in Russian, English summary).

Lavrov, A. S. 1978: Some types of kames in the southern marginal zone of the Barents ice sheet. *Stroyenie i formirovanie kamov*, 53–61. Tallinn (in Russian, English summary).

Lavrov, A. S., Nikiforova, L. D. & Potapenko, L. M. 1986: The dynamics of the Pleistocene ice sheets, vegetation and climate in the northeast of the European USSR. *In* Sidnev, A. V. & Nemkova, V. K. (eds.): *Novye materialy po paleogeografii i stratigrafii pleistotsena*, 69–78. Bashkir Branch Acad. Scis. USSR, Ufa (in Russian).

Lavrov, A. S. & Potapenko, L. M. 1989: A comparative characteristic of Late Pleistocene glacial features and terraces in the north of the Pechora Lowland and West Siberia. *In* Velichko A. A. Gurtovaya Ye. & Faustova M. A. (eds.): *Paleoklimaty i oledeneniya v pleistotsene*, 205–211. Nauka, Moscow (in Russian).

Lavrov, A. S., Potapenko, L. M., Zaitsev, A. P., Korolkova, Z. G., Meshalkin, K. A. & Nikiforova, L. D. 1991: Map of the Quaternary deposits, scale 1:1 000 000, sheet Q-38,39 (Mezen). *The National Geological Map of the USSR, new series*. Mingeo SSSR, Leningrad.

Lavrushin, Yu. A., Chistyakova, I. A., Gaidamanchuk, A. S., Golubev, Yu. K. & Vasilyev, V. P. 1989: The structure and composition of the glacial paleoshelf sediments in Bolshezemelskaya Tundra. *Litologia kainozoiskikh otlozheniy* 3–51. Geol. Inst. Acad. Scis. USSR, Moscow (in Russian).

Levkov, E. A. 1980: *Glaciotectonics (Glatsiotektonika)*. 280 pp. Nauka i tekhnika, Minsk (in Russian).

Mangerud, J., Svendsen, J. I. & Astakhov, V. 1999: The age and extent of the Barents and Kara Sea ice sheets in Northern Russia. *Boreas 28*, 46–80.

Popov, A. I. 1962: Cover silts and polygonal relief of Bolshezemelskaya Tundra. *In* Popov A. I. (ed.): *Voprosy geographicheskogo merzlotovedenia i periglatsialnoi morfologii*, 109–130. Moscow University (in Russian).

Punkari, M. 1995: Glacial flow systems in the zone of confluence between the Scandinavian and Novaya Zemlya ice sheets. *Quaternary Science Reviews 14*, 589–603.

Svendsen, J. I., Astakhov, V. I., Demidov, I., Bolshiyanov, D. Y., Dowdeswell, J. A., Gataullin, V., Hjort, C., Hubberten, H. W., Larsen, E., Mangerud, J., Melles, M., Möller, P., Saarnisto, M. & Siegert, M. J. 1999: Maximum extent of the Eurasian ice sheets in the Barents and Kara Sea region during the Weichselian. *Boreas 28*, 234–242.

Tarakanov, L. V. 1973: To the issue of origin of the relief of the Vaigach Island. *Geomorfologia 4*, 85–91 (in Russian).

Tveranger, J., Astakhov, V. & Mangerud, J. 1995: The margin of the last Barents–Kara ice sheet at Markhida, Northern Russia. *Quaternary Research 44*, 328–340.

Tveranger, J., Astakhov, V. I., Mangerud, J. & Svendsen, J. I. 1998: Signatures of the last shelf-centered glaciation at a key section in the Pechora Basin, Arctic Russia. *Journal of Quaternary Science 13*, 189–203.

Tveranger, J., Astakhov, V., Mangerud, J. & Svendsen, J. I. 1999: Surface form of the last Kara Ice Sheets as inferred from its southwestern marginal features. *Boreas 28*, 81–91.

Velichko, A. A., Kononov, Yu. M. & Faustova, M. A. 1997: The last glaciation of Earth: size and volume of ice-sheets. *Quaternary International 41/42*, 43–51.

Voronov, P. S. 1951: New data on the glaciation and Quaternary sediments of the central Pai-Hoi. *Sbornik statei po geologii Arktiki 2*, 84–92. Inst. Geol. of Arctic, Leningrad (in Russian).

Yakovlev, S. A. 1956: The fundamentals of the Quaternary geology of the Russian Plain. *Trudy VSEGEI 17*, 314 pp. Moscow (in Russian).

Yershov, E. D. (ed.) 1988: *The geocryology of the USSR. European territory (Geokriologia SSSR. Yevropeiskaya territoria)*. 358 pp. Nedra, Moscow (in Russian).

Yudkevich, A. I. & Simonov, A. N. 1976: The stratigraphy of the Pliocene and Pleistocene of the Pechora catchment area. *In* Yakhimovich V. L.(ed.): *Voprosy stratigrafii i korrelatsii pliotsenovykh i pleistotsenovykh otlozheniy severnoi i yuzhnoi chastei Preduralya*, 142–164. Ufa (in Russian).